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SOME PROBLEMS OF CHEMICAL MINERALOGY*†

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INTRODUCTION

The retiring president of the Mineralogical Society is customarily called upon to deliver an address, and although no restrictions are laid upon him, it is generally expected that from the depths of his experience he will deliver words of wisdom to the assembled multitude on some phase of mineralogy in which he is well qualified. Having heard quite a few such addresses and now having gone through the experience of preparing one, I am in favor of the practice, mostly because I think it is extremely valuable to the retiring president to be forced in this way to reflect on what he has done and the relation of his accomplishment to the broad field of mineralogy.

It must admit that when I began to think of what to say here today I was rather perplexed. My work for the past eight years has been largely administrative, so that if this talk were to reflect my recent work, it would have to deal with subjects such as whether papers can be moved more efficiently from right to left or vice versa, and whether when planning a budget one should allow 5% or 10% as a safe contingency fund. This might have been instructive, but probably not interesting. It occurred to me that as an alternative, I might read to you some choice excerpts from letters received from persons whose descriptions of new minerals I have abstracted in the *American Mineralogist*. This would have been interesting, but probably not instructive.

Fortunately it has also been necessary for me as a member of Earl Ingerson's staff in the Geochemistry and Petrology Branch of the U. S. Geological Survey to do a good deal of thinking about the functions of a laboratory whose purpose is the application of chemical, spectrographic, mineralogical, *x*-ray, and petrographic techniques to the solution of geological problems. As one result of such thinking, I have accumulated over a period of years notes on research problems that seemed to me to be worthy of attention, and I propose to discuss some of these problems today. This talk might thus have been entitled "Some Problems I

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* Publication authorized by the Director, U. S. Geological Survey.

† Address of the retiring President of the Mineralogical Society of America given at the thirty-third meeting of the Society in Boston, Nov. 14, 1952.

Hope I Have Time to Tackle Some Day." The point of view is primarily that of a chemist. Perhaps I should apologize for restricting myself to problems of chemical mineralogy—my excuse is not lack of appreciation of other phases of mineralogy, but merely my ignorance of them. I am afraid that I have the distinction of being the only President of this Society whose training was so concentrated in the field of chemistry that according to the requirements set forth by the U. S. Civil Service Commission, I am not eligible for appointment to the position even of junior mineralogist in the government service. So I shall speak as a chemist—perhaps we can say as a geochemist—about several problems that are not too closely related to one another except that all of them need attention by chemists.

STABILITY RELATIONS OF POLYMORPHS

The first such problem to be discussed is the problem of the stability relations and the conditions of formation of polymorphous substances. These substances are familiar to all mineralogists; a few of the best-known examples are listed in Table 1. These compounds have been the subject of researches for many decades—crystallographic, chemical, physical, and, more recently, *x*-ray studies. They have been and are of special interest to the economic geologist in that they offer the possibility of furnishing clues as to the conditions of formation of mineral deposits. What do we really know of the stability relations and conditions of formation of the polymorphs listed in Table 1? Not nearly as much as most of us would like.

TABLE 1. SOME POLYMORPHOUS NATURALLY OCCURRING SUBSTANCES

ZnS	Sphalerite—Wurtzite
HgS	Cinnabar—Metacinnabar
FeS_2	Pyrite—Marcasite
Fe_2O_3	Hematite—Maghemite
As_2O_3	Arsenolite—Claudetite
Sb_2O_3	Senarmontite—Valentinite
TiO_2	Rutile—Anatase—Brookite
SiO_2	Quartz—Tridymite—Cristobalite
$CaCO_3$	Calcite—Aragonite
Al_2SiO_5	Kyanite—Andalusite—Sillimanite
$MgSiO_3$	Enstatite—Clinoenstatite

Nearly 40 years ago, the Geophysical Laboratory published a series of papers on the stability relations and conditions of formation of the silica minerals (Fenner, 1913), the sulfides of iron, zinc, and mercury (Allen *et al.*, 1912, 1914), and calcium carbonate (Johnston, Merwin, and

Williamson, 1916). In these papers, the effects of temperature, acidity, and composition were carefully considered. Unfortunately the results have been misinterpreted by some later workers; in an understandable attempt at simplification, too much stress has been laid on the effects of temperature and acidity, and not enough on the effects of variation in composition. Loose statements like the following have been made all too frequently:

> The presence of wurtzite indicates that the ore-forming solution was acid,

or

> Cinnabar is precipitated only from alkaline solutions.

Such statements *may* be correct, but in the words of the modern psalmist, "It ain't necessarily so." Such statements neglect the effect of composition.

The researchers at the Geophysical Laboratory clearly recognized the importance of the effect of foreign ions on the stability of polymorphous forms. This is shown by the graph on page 348 of the 1912 paper of Allen, Crenshaw, and Merwin for the pair sphalerite-wurtzite. Even more striking effects might be expected when one of the polymorphs is the unstable form at all temperatures when pure. Johnston, Merwin, and Williamson (1916) discussed this in some detail, and after citing experimental work going back 100 years remarked (p. 500), "It is possible that aragonite containing other material in solid solution may not be unstable with respect to calcite," and (p. 509) "Now the solubility of pure aragonite is not much greater than that of calcite, and this difference would be diminished by the presence of other materials in solid solution in the aragonite; it is conceivable, therefore, that such impure aragonite should be really stable in contact with the natural waters in which it formed. On this basis, then, it is plausible (though not yet established) that natural aragonite containing other substances in solid solution should be practically stable under certain conditions." Today, 36 years later, it is still plausible, and still not yet established for $CaCO_3$.

Allen and his coworkers showed that metacinnabar was metastable with respect to cinnabar at all temperatures. Recently Rittner and Schulman (1943) studied the effects of cadmium on the transition metacinnabar-cinnabar in synthetic solid solutions. The data summarized in Table 2 show that even a very small amount of cadmium has a very marked effect on the transition, and it is possible that metacinnabar might be stable if enough cadmium were present in solid solution. Naturally occurring metacinnabar commonly contains appreciable amounts of zinc and iron, and spectrographic analyses made in the laboratory of the Geological Survey by Cyrus Feldman have shown that germanium is present in most cinnabar and metacinnabar. On this basis, it is plausible (though not yet established) that natural metacinnabar containing other substances in solid solution should be practically stable under certain conditions.

TABLE 2. HgS-CdS SOLID SOLUTIONS TREATED WITH $(NH_4)_2S$ SOLUTION

Wt. % Cd	Time of Treatment	Result
0	2 days	100% cinnabar
0.006	3 days	100% cinnabar
0.13	14 days	100% cinnabar
0.32	10 weeks	Cinnabar, little metacinnabar
0.64	10 weeks	Metacinnabar, cinnabar
1.3	10 weeks	Metacinnabar

Data of Rittner and Schulman, 1943.

Recent analyses in our laboratory show that rutile from the alkalic rocks at Magnet Cove, Arkansas, contains 1.2% niobium, whereas brookite from the same locality contains 2.4% niobium. Is this accidental, or cause and effect or, as W. T. Pecora suggests, shrewd sampling on the part of the field geologist? Unfortunately, comparatively little experimental work has been done in the past 25 years. Some years ago, a series of papers by M. J. Buerger and M. C. Bloom of the Massachusetts Institute of Technology was devoted to consideration of polymorphism (Buerger, 1934, 1935, 1936); Buerger and Bloom (1937); Bloom and Buerger (1937); Bloom (1939). These papers discussed the effect of variation in composition; it was shown experimentally for Sb_2O_3 that the metastable form persisted if certain impurities were present in the compounds.* A similar explanation was offered for the persistence of tridymite and cristobalite at low temperatures (Buerger, 1935, Fig. 2, p. 190). The explanation seems reasonable for such open framework structures; it is strengthened by the discovery of a variety of tridymite, named christensenite, that contains about 5% of $NaAlSiO_4$ in solid solution (Barth and Kvalheim, 1944). Recent x-ray studies strongly indicate that the presence of impurities has considerable effect in determining which polymorph formed in syntheses of Bi_2O_3 (Sillén, 1938) and of silicon carbide (Lundqvist, 1948).

Thus we have good evidence that both direct atom-for-atom substitution, as in wurtzite or metacinnabar, and the interstitial type of foreign ions in open structures, as in Sb_2O_3, may have profound influence on the stability relations of polymorphous substances. I should like to cite just one more pair of minerals that deserves study—the dimorphs of Fe_2O_3, hematite and maghemite. Unpublished studies made over a period of years by my colleague Charles Milton have shown that maghemite, gamma-Fe_2O_3, occurs far more commonly than is generally supposed. It is known to be unstable with respect to hematite. Why then is it formed in Nature, and why does it persist? We do not know. If I may be permitted to speculate, is it possible that the presence of non-oxidizable ions such as magnesium or nickel, perhaps filling some of the vacant iron positions in the gamma-Fe_2O_3 structure, causes the formation of maghemite rather than of hematite? Plausibility is lent to this speculation by the observation that maghemite is abundant in the oxidized zones of a nickel-iron sulfide deposit. We hope that it may be possible to isolate and analyze pure samples of maghemite and to study the course of oxidation of synthetic magnetites containing known amounts of magnesium and nickel.

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* A different explanation was recently advanced by Frueh (1952), who believes that a hydrogen bond is present in valentinite.

In view of the very great interest in trying to use polymorphous forms as guides to learning the conditions of formation of ore deposits, it is discouraging to note how little experimental work has been done in this field for the past 20 or 30 years. We need new analyses, chemical and spectrographic, for major and minor elements on samples of known environment; we need new studies on mineral synthesis, dry and hydrothermal. Here is plenty of work for the chemist.

RELATION OF COMPOSITION TO PARAGENESIS

So much then for the effect of the environment on which form of a polymorphous compound may be precipitated. Let us turn to another familiar theme—the variation of composition of minerals as related to the environment in which they were formed. It is gratifying that this theme has become so familiar. I should like to recall to you two studies of this sort, and then to consider other work that might profitably be undertaken.

The first mineral I shall discuss is the uncommon mineral aphthitalite, a rhombohedral solid solution of potassium and sodium sulfates (note that K_2SO_4 and Na_2SO_4 are orthorhombic). The mineral is of unusual interest because it has two very different modes of origin: crystallization from saline solutions at temperatures below 100° C., and formation as a product of volcanic and fumarolic activity at relatively high temperatures. Figure 1 shows analyses of 10 samples from potassium salt deposits (Germany, Poland, California, and New Mexico); the range of concentration is small, with most samples having K:Na about 3:1. The

25 analyses shown of aphthitalite of volcanic origin (from Vesuvius, Etna, and Kilauea) show a much wider range of composition, the K:Na ratio ranging from 3:1 to 1:2; they also show the presence of appreciable amounts of Pb, Cu, and Ca in solid solution. The system K_2SO_4-Na_2SO_4-H_2O has been studied (Hamid, 1926; Druzhinin, 1938) and it has been found that the range in composition of the rhombohedral solid solutions that are stable is very small. The system K_2SO_4-Na_2SO_4 and K_2SO_4-Na_2SO_4-$PbSO_4$ (Perrier and Bellanca, 1940), K_2SO_4-Na_2SO_4-$CuSO_4$ (Bellanca and Carapezza, 1951), and K_2SO_4-Na_2SO_4-$CaSO_4$ (Bellanca, 1942) have all been studied. The results are in good agreement with the observations on natural material; the range of solid solutions in the anhydrous systems at melting temperatures is much more extensive than in the system with water.

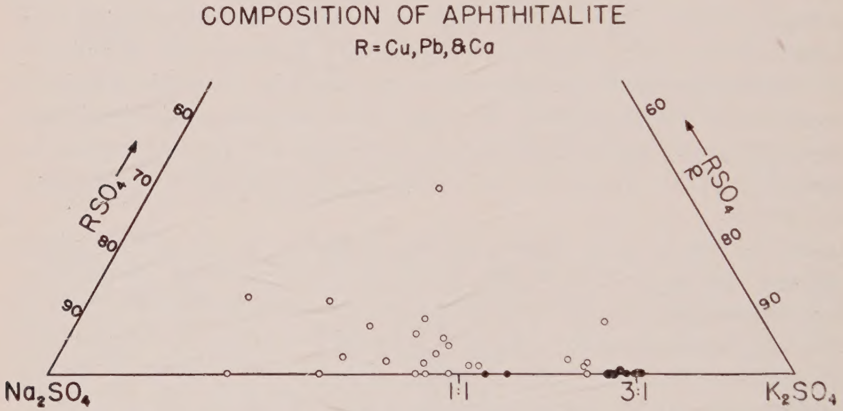

FIG. 1. Composition of aphthitalite.

Solid circles = samples from salt deposits
Open circles = samples from volcanoes or fumaroles

I have discussed this uncommon mineral because the relations are so simple that they should be better known, especially to teachers. Let us turn now to more common minerals. It is surprising how little information is readily available even for the commonest minerals as to how their composition varies with their geological environment. There are some minerals, including some of the most common ones, for which we know that we have very little information. For example, this was true of the mineral chromite until about 10 years ago—partly because its analysis is extremely difficult. We now have two fine sets of analyses of chromite, one set of chromites from the Bushveld complex (van der Walt, 1941), the other of chromites studied by the U. S. Geological Survey from many ore deposits in the Western Hemisphere (Stevens, 1944). The latter shows that the composition of chromite varies much more than has generally been supposed. His results were interpreted by Thayer (1946) from the standpoint of the geological environment. Thayer showed that when Stevens' analyses are grouped on a regional basis, considerable order appears in what seemed to be random variation in composition and that this order reflects the effect of the geological environment. Thus the alumina-rich chromites of Cuba are closely associated with troctolites and gabbros. Petrographic evidence shows that the chromite has reacted with the gabbroic magma with the formation of anorthite at the expense of part of the alumina content of the chromite, and this explains why these chromites show a rather wide range in Al_2O_3 content. In contrast, the chromites from the feldspar-free peridotites of the Pacific coast are low in Al_2O_3. We have here just the beginnings of deciphering a complex problem; the U. S. Geological Survey has under way a detailed study of the chromites of the Stillwater complex in Montana, which we hope will help to fill some of the gaps in our knowledge.

These are examples of what has been done; we need further work on many other minerals. There are some regional studies, as for example, those by Professor Larsen and his coworkers on the variation in composition of such minerals as biotite and amphibole in the San Juan district, Colorado (Larsen et al., 1936, 1937, 1938) and in the southern California batholith (Larsen and Draisin, 1950). For some minerals, it is hard to say whether the data are adequate. Summaries have been published for a few, as for example for the garnet group (Wright, 1938). Similar compilations for other minerals are badly needed. For example, the existing analyses of the amphiboles and the micas need to be assembled and studied critically, and every relationship of composition to physical and optical properties and to geological environment ought to be squeezed out. Not only would this help greatly in systematizing what we know, but it would certainly show us where the gaps are that need to be filled. Probably no one will disagree as to the need; just as probably, few if any will be willing to undertake such a task of compiling. We are all prone to take compilations for granted, to be extremely critical of the way they are done, and to consider them somewhat unworthy of our own great talents. This is really too bad, because there is little doubt that a good critical review saves much more time than it takes and points out valuable researches that ought to be done. As an example, a critical study of the 50 most recent analyses of biotite made in 1941 by J. J. Fahey, R. E. Stevens, and me showed that there were glaring deficiencies in the published determinations of Li_2O, F, and H_2O, and alerted our laboratory to the need for better methods of analysis for these constitutents of micas.

One other mineral ought to be mentioned. We hope soon to begin work on the mineral magnetite. Examination of the literature shows relatively few analyses of pure magnetite, yet even these few are very suggestive. For example, there are just a few analyses of magnetite from diabases and Tertiary lavas and every one of these is very high in MgO content—in fact two of them are actually magnesioferrite rather than magnetite. There are also a few analyses of magnetites from nepheline syenites and phonolites; each shows a high content of MgO and Al_2O_3 and an appreciable content of MnO as well. We have been collecting and purifying samples of magnetite of various origins and we hope to have something to report in the future.

By this time, each of you has probably thought of a dozen other minerals that ought to be studied in this way; there ought to be no shortage of work for the chemists!

Distribution of Minor Elements among Minerals of a Rock

I shall go on now to another, closely related problem—that of the minor elements in rocks and their distribution among the constituent minerals of the rocks. Twenty-five years ago, Professor Papish of Cornell University lectured to the Geological Society of America on the possibilities of the spectrographic method of analysis, and V. M. Goldschmidt and the Noddacks used the method in the early 1930's to increase our knowledge of the abundance and distribution of many of the rarer elements. But it has taken 20 years of patient work to develop the method and to standardize the results, and even now only a handful of laboratories in North America are applying the spectrograph to the solution of geological and mineralogical problems. The developments in this field during the past few years, plus the rapid advances in the use of colorimetric methods for trace amounts that have been led by Professor Sandell at Minnesota, promise to bury us under a mass of data.

Again I should like to give a few examples of research that have been done and to mention some additional problems that ought to be studied. To date there have been very few studies in which the distribution of the elements among the minerals of the rocks has been determined. One such study, of exceptional interest, is that by C. S. Ross, M. D. Foster, and A. T. Myers (1953) on the minerals of dunites and of olivine-rich inclusions in basalt. They determined the major and minor elements present in a series of olivines, orthopyroxenes, clinopyroxenes, and spinels separated from rocks from many localities. The distribution of elements such as Ni, Co, Cr, and Mn among these minerals is of particular importance. Two other reports should be mentioned: the study of

Caledonian rocks of Scotland by Nockolds and Mitchell (1948), and that of the igneous rocks of the Skaergaard intrusive of eastern Greenland (Wager and Mitchell, 1943, 1951). The mineralogical and geochemical implications of their data can scarcely even be touched upon here; only a few points can be mentioned. For example, it is interesting to compare the data of Nockolds and Mitchell on granites, granodiorites, and similar rocks with those of Ross and coworkers on ultramafic rocks. Nockolds and Mitchell found chromium to be highly concentrated in monoclinic pyroxene, with much lower concentrations in hornblende and biotite; they found lithium to appear almost exclusively in biotite, and vanadium to appear in highest concentrations in the micas, with appreciable but lesser amounts in pyroxenes and hornblende.

The Skaergaard intrusive is the only highly differentiated igneous rock mass for which we have detailed work on major and minor constituents. One must be cautious in drawing generalizations, since the Skaergaard magma was somewhat unusual in composition, particularly in its high content of iron and low content of potassium. The data of Wager and Mitchell show that chromium and nickel are largely precipitated, like magnesium, in the first fractions to crystallize, whereas cobalt and vanadium come out over a much wider range of concentration. If we now consider further the distribution of these elements in the individual minerals, we begin to get a picture of what happens during differentiation and we can compare the data for these gabbroic rocks with those for granodiorites and ultramafics. The nickel, which is largely in the first fractions, is almost entirely in the olivine; cobalt precipitates over a wider range of composition and is distributed between the olivine and pyroxene; and copper precipitates still later and is present in both ilmenite and magnetite as well as in pyroxene. Vanadium behaves very much like copper. In the earlier stages, vanadium is present almost entirely in the pyroxene; as crystallization proceeds, more and more vanadium is present in the magnetite, up to 0.5% in magnetite of some of the gabbros.

Several applications to problems of economic geology are apparent. The data show that most of the copper content of these rocks is present in silicate minerals, and it is only at a late stage of differentiation that copper-bearing sulfides appear. Another point is this—can the results explain the fact that deposits of titaniferous magnetite the world over that occur in close association with gabbros and anorthosites invariably contain high concentrations of vanadium? Wager and Mitchell have pointed out that crystallization differentiation accounts for concentrations of 0.5% or thereabouts in the magnetites of the Skaergaard intrusive, but that many titaniferous magnetites from other localities contain much higher concentrations of vanadium. We badly need similar studies of other intrusive masses before we can assume that the mechanism of differentiation offers an adequate explanation.

These few examples are perhaps enough to indicate that the possible applications of further research along these lines to mineralogy, economic geology, and petrology are almost unlimited. I shall cite just two other major problems.

The first is the problem of the composition, mineralogy, and mode of origin of the marine black shales. Comparatively little work was done on these until recent years, but unpublished work by the U. S. Geological Survey plus several published papers (Westergård, 1944; Goldschmidt *et al.*, 1948; Schneiderhöhn *et al.*, 1949; McKelvey and Nelson, 1950; Leutwein, 1951) give just enough information to show how complex and difficult the problem is. These rocks contain, on the average, more vanadium, more chromium, more copper, and more molybdenum—not to mention more uranium—than most igneous rocks. Sad to say, we know very little of the form in which these elements are present in the black shales. We need to separate the minerals and to find out where these elements are concealed before we can hope to solve the problem of the mode of formation of these rocks.

The second problem to which this type of research needs to be applied is that hardy perennial—the origin of alkalic rocks such as the nepheline syenites. Forty years ago, C. H. Smyth (1913) pointed out that any explanation of the origin of these rocks had to account for the unusual concentration in them of elements such as the rare earths, titanium, zirconium, and thorium. In spite of that, few of the multitude of papers on the subject have even mentioned this aspect. We do not have enough determinations of the minor elements in alkalic rocks, and still fewer of their distribution among the minerals of such rocks. Our laboratory has begun such studies, under the direction of W. T. Pecora, not only because of the petrological problems, but also because it seems likely that these rocks may become important sources of such elements as titanium, the rare earths, and niobium.

Application of Mineralogical Thinking to Problems of Extractive Metallurgy

Finally, I should like to consider briefly a fourth type of problem—the use of the mineralogical approach to problems of extractive metallurgy. This, of course, is not new. It is not a very profound idea that when one attempts to separate two elements occurring together in an ore, it helps to know in what minerals the elements are present. Unfortunately there are still too many metallurgists and chemists to whom this idea has not yet penetrated. I should like to quote to you three statements made in 1944 and 1945 about three different manganese oxide ores:

1. "Gravity concentration and flotation failed to reduce the BaO content, indicating that the BaO is very intimately associated with the manganese." This statement led us to investigate the ore, which was found in the U. S. Geological Survey laboratory to consist chiefly of the barium manganese oxide mineral psilomelane.

2. "Jigging, high-intensity magnetic separation, and flotation failed to remove the contamination by zinc." Since zinc-bearing manganese oxides are not very common, this ore was also investigated. It contains as a major constituent the mineral hetaerolite, $ZnMn_2O_4$.

3. "Your suggestion that the mineral is a lead-bearing hollandite is of great interest, since a study of practical means of reducing the lead content of the concentrates has led us to the conclusion that most of the lead was not amenable to separation by customary mechanical or chemical means."

It is saddening to think of the wasted effort in each of these instances in the attempt to separate elements that were combined in a single mineral. The point is not that mineralogists should feel smug and superior; on the contrary, the moral I should like to draw is that these are illustrations of *our* failure to educate metallurgists and chemists to recognize the proper role of the mineralogist and the need for his services instead of tackling such problems by empirical methods. If mineralogists did not hold themselves aloof from these practical problems, they would perhaps create more widespread recognition of what they can do.

Let us look at the more positive side of the mineralogical aspects of problems of extractive metallurgy. Industrial developments of the past few years have made it imperative to find new sources of some of the rarer elements. One such is niobium (columbium), and I should like to mention briefly work in our laboratory aimed at finding new sources of this element. Details are being published elsewhere (Fleischer *et al.*, 1952).

From general geochemical considerations and from the data in the literature, it appeared that the highest concentrations of niobium are in granitic pegmatites and in alkalic rocks; further that niobium should be closely associated with the elements titanium, zirconium, and tungsten. With these generalizations as guides, we have determined the niobium content of many titanium minerals, and we have found that titanium minerals associated with gabbros and anorthosites have very low niobium contents, whereas rutile, perovskite, and ilmenite from alkalic rocks are high in niobium and constitute potential sources of that element.

Another point of attack started from data that showed (Murata and Gordon, 1952) that nepheline syenites from Arkansas contained appreciable amounts of niobium, and further that Arkansas bauxites formed by the weathering of these alkalic rocks contained even more niobium, about 0.05% Nb on the average—about a pound per ton. This is perhaps not very impressive, but calculation shows that the Arkansas bauxite processed annually contains about nine-tenths as much niobium as the total annual world production.

Figure 2 gives a simplified flow sheet of the treatment of bauxite with the niobium contents of the various products. The niobium is concentrated in the insoluble fractions, of which the brown mud forms the major part. We have not yet attempted mineralogical fractionation of this extremely fine grained material. Fractionation of the black sand by means of heavy liquids and magnetic separations gave the data of Table 3. It will be noted that a fraction was obtained that contained 0.66% Nb and 46% TiO_2, one that may therefore be of commercial interest.

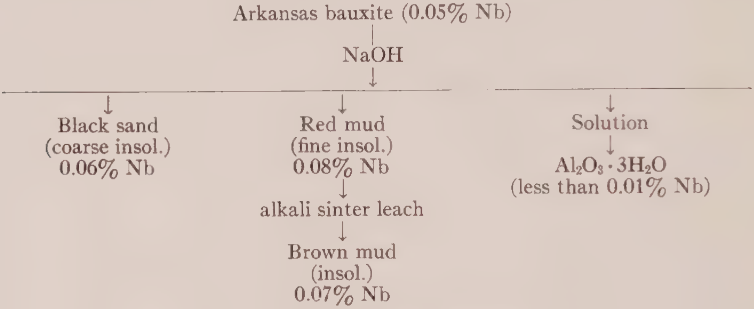

FIG. 2. Nb IN BAUXITE TREATMENT.

This is a simple illustration of why it is not enough to know the concentration of an element in an ore and why we also need to know the element's concentration in the different minerals of the ore. The general trend in extractive metallurgy in recent years has been toward the utilization of ever lower grades of ore and also toward the recovery of

TABLE 3. FRACTIONATION OF BLACK SAND FROM ARKANSAS BAUXITE

Material	Percent of Total	Percent Nb	Percent TiO_2
Black sand	100	0.064	not detd.
Float at 2.9 (calcite, quartz, etc.)	14.5	not detd.	not detd.
Magnetite	51.5	0.025	5.2
Mainly limonite	14.8	0.04	not detd.
Ilmenite+limonite	15.9	0.14	not detd.
Ilmenite	2.8	0.66	46
Zircon+sphene+rutile	0.5	0.45	not detd.

Mineralogical separations by Marie L. Lindberg; spectrographic analyses by Janet D. Fletcher.

more and more of the minor elements. I believe that mineralogists can and should play a very important role in the future development of this field.

CONCLUSION

I have discussed some of the chemical problems of mineralogy that deserve more attention. They require more than routine analyses by chemists; they demand the best thinking of chemists of research caliber. It is the duty of mineralogists and geologists to arouse the interest of chemists in such problems and to enroll them in full partnership in these researches.

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CRYSTAL GEOMETRY OF SOME ALKALI SILICATES

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Abstract

α-$Na_2Si_2O_5$ (the high-temperature form) is orthorhombic *Pnam*, pseudo-*Pnab*. $a=15.45\pm0.03$, $b=4.909\pm0.005$, $c=6.428\pm0.015$ Å. Z=4. Density (g/cm^3) $=2.47\pm0.03$ (obs.), 2.48 (calc.). Thick tabular {100}. Cleavage {100}.

β-$Na_2Si_2O_5$ is monoclinic $P2_1/a$. $a=12.307\pm0.005$, $b=4.849\pm0.003$, $c=8.124\pm0.004$ Å, $\beta=104.12\pm0.005°$. Z=4. Density (g/cm^3) $=2.57\pm0.10$ (obs.), 2.542 (calc.). Thin tabular {100}, elongated [010]. Cleavages {100} and {001}. Dual twinning, twin plane and composition plane (100).

$Li_2Si_2O_5$ is orthorhombic, diffraction aspect *Ccc**. $a=5.80\pm0.02$, $b=14.66\pm0.05$, $c=4.806\pm0.015$ Å. Z=4. Density (g/cm^3) $=2.454$ (lit., at 25° C.), 2.438 (calc.).

Li_2SiO_3 is orthorhombic, diffraction aspect *Cc***, probable space group *Ccm2*. $a=5.43\pm0.02$, $b=9.41\pm0.03$, $c=4.660\pm0.015$ Å. Z=4. Density (g/cm^3) $=2.520$ (lit., at 25° C.), 2.50 (calc.).

Introduction

In the course of a survey of soluble silicates, we examined crystals of two polymorphic forms of sodium disilicate $Na_2Si_2O_5$, lithium disilicate $Li_2Si_2O_5$, and lithium metasilicate Li_2SiO_3. All *x*-ray photographs were taken with CuKα radiation ($\lambda=1.5418$ Å). The sodium compounds were prepared by Dr. G. W. Morey, the lithium compounds by Dr. F. C. Kracek.

Sodium Disilicates

The phase diagram Na_2O-Si_2O has been described by Morey and Bowen (1924) and by Kracek (1930). The melting point of $Na_2Si_2O_5$ is $874\pm1°$. Kracek has obtained three polymorphic forms, distinguished by their powder patterns, on which he reports as follows (private communication). The lowest-temperature form is obtained only on crystallization from the glass between 500 and 580°. It has not yielded any single crystals and its powder pattern has not been indexed. It appears to be a metastable modification. Hydrothermally, between 480 and 670°, another form (β) crystallizes. This β form is also obtained from the glass between 580 and 670°. Above 670° the high-temperature form (α) appears. Both α and β forms are somewhat hygroscopic. The transition from β to α is sluggish; at 695° it takes several days. Attempts to transform the α to the β-form have been unsuccessful. Further details on occurrence and properties will be published in a forthcoming paper from the Geophysical Laboratory.

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(1) α-$Na_2Si_2O_5$

α-Sodium disilicate is orthorhombic, apparently holohedral. Single crystals measuring 3 to 4 mm. in maximum dimension have been found. They form milky white plates, thick to very thick tabular on {100}, slightly elongated [010]. They possess one very good and easy cleavage {100}, which gives mediocre reflections. All other reflections are very poor. The plate is beveled by a rhombic prism, which is symbolized {201} in order to conform with the *x*-ray results (see below). A small {010} face is present, but the pinacoid {001} is missing. The measured angle $(100):(201) = 49\frac{3}{4}°$ yields the crude axial ratio $c/a = \frac{1}{2} \tan 40\frac{1}{4}° = 0.42$, which agrees with $c/a = 0.416$ obtained by *x*-rays.

A sample of fibrous sodium disilicate (not commercially available) was given to us by the Philadelphia Quartz Company. It proved to be the α-form.

The optical data given by Bowen (in Kracek, 1930*a*) are as follows: $n_\alpha = 1.497$, $n_\gamma = 1.508$, biaxial negative, $2V = 50$–$55°$, two cleavages parallel to $\beta\gamma$ and $\gamma\alpha$.

Weissenberg photographs were taken about the three axes. The *x*-ray extinctions indicate the diffraction aspect *Pna**, which is compatible with two space groups: *Pna*2 and *Pnam*. Inasmuch as the goniometric measurements seem to show that both the *a* and the *b* axes are 2-fold axes, the crystal is most likely holohedral with space group *Pnam*. In addition to the systematic space-group extinctions, near-extinctions are also observed—the *hk*0 reflections with *k* odd are very weak.[1] They indicate a pseudo *b* glide plane parallel to (001) and a pseudo space group *Pnab*.

The cell dimensions were refined with the help of a powder pattern (taken by Dr. H. Yoder on the Philips back-reflection *x*-ray spectrometer). They are: $a = 15.45 \pm 0.03$, $b = 4.909 \pm 0.005$, $c = 6.428 \pm 0.015$ Å. This choice of axes is made to bring out the relationship with the monoclinic form, so that the cleavage is symbolized {100} and the *b* unit length is nearly the same in both forms. The density, based on three determinations on the Berman balance, with crystals weighing 4 to 5 mg, is 2.47 ± 0.03 g/cm^3. With 4 formula units per cell, the calculated density is 2.48 g/cm^3.

Fiber diagrams obtained with the material from the Philadelphia

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[1] Only 210, 13.1.0, and 17.1.0 are observed at all. They were obtained on two different photographs, taken with different samples (*b*-axis first layer), and are therefore considered reliable reflections.

Quartz Company show that the *b* axis is the fiber axis. The reflections are drawn out along arcs which subtend an angle of 13° at the center of the film. It follows that the fibrils are not perfectly parallel to the fiber axis but may deviate from it as much as $6\frac{1}{2}$°. The diffracted arcs are sharp, indicating a fibril thickness of at least 1000 Å. The fibrous material is as hygroscopic as the single crystals.

(2) β-$Na_2Si_2O_5$

β-Sodium disilicate is monoclinic, apparently holohedral. Some single crystals measure up to 2 mm. in maximum dimension, most of them less than one mm. They form colorless pseudo-hexagonal plates or laths, thin tabular on {100} and elongated [010]. They show two cleavages: {100}, eminent, nearly micaceous, and very easy; {001}, good and fairly easy, indicated by [010] striations on {100}.

Except for the faces of the pinacoid {100}, which give excellent reflections, the crystals are of poor quality for optical goniometry. Three relatively satisfactory angles, namely, (100):(001) = 75°45′, (100):(201) = 31°25′, $(011):(01\bar{1})$ = 63°20′, were obtained on a two-circle instrument. They yield the following crude axial elements:

$$a:b:c = 2.49_3:1:1.67_1; \qquad \beta = 104°15'.$$

The forms observed are: {100} dominant; {001}, {011}, {201}, and $\{\bar{2}01\}$, common; {101}, {401}, {601}, uncertain. The occurrence of {001} as a subordinate form, with {200} as the habit-controlling form, is anomalous. The morphological development is otherwise consistent the space group $P2_1/a$ determined by *x*-rays (see below). The latter requires the point group to be $2/m$, thus confirming the holohedry.

Dual twinning is universal, with twin plane (100). It is twinning by reticular pseudo-merohedry, due to the quasi-perpendicularity of [310] and (100). The twin index is 3, its obliquity 1°21′. The composition surface is planar and parallel to the twin plane, as required by theory. Owing to the extreme thinness of the plates, such a contact twin simulates crystals in parallel orientation. The two individuals, lying on (100), extinguish of course simultaneously between crossed nicols, so that this kind of twinning may easily pass unnoticed, even optically.

The refractive indices were re-measured by Dr. Jewel Glass: $n_\gamma = 1.515$ lies along [010]; the second index (1.510) measured in the (100) plane and the index (1.500) measured perpendicular to it, do not differ much from the principal indices n_β and n_α respectively, as the interference figure obtained in convergent light shows the acute bisectrix nearly perpendicular to (100). The optical character is negative (2E = 105°, in Morey, 1914).

Weissenberg photographs were taken about [100] and [010]. The space

TABLE 1

β-$Na_2Si_2O_5$

$a=12.307\pm0.005$ $b=4.849\pm0.003$ $c=8.124\pm0.004$ Å

$\beta=104.12\pm0.005°$

Qual.*	2θ obs. Cu K_α	*d* obs.	Q obs.	Q calc.	2θ calc. Cu K_α	*hkl*
1 *R*	11.24	7.871	.01614	.01611	11.23	001
9 *R*	14.84	5.970	.02806	.02808	14.84	200
3 *R*	16.27	5.448	.03369	.03381	16.30	$20\bar{1}$
				.04955	19.76	110
9 *R*	20.74	4.283	.05452	.05457	20.75	201
4 *R*	21.53	4.127	.05871	.05864	21.52	011
				.06047	21.86	$11\bar{1}$
10 *R*	22.57	3.939	.06444	.06444	22.57	002
3	23.65	3.762	.07066	.07061	23.64	210
				.07085	23.68	111
				.07177	23.84	$20\bar{2}$
8 *R*	24.58	3.622	.07624	.07634	24.59	$21\bar{1}$
				.09710	27.80	211
1 *R*	28.71	3.109	.10344	.10361	28.73	$11\bar{2}$
				.10571	29.03	310
				.10625	29.11	$31\bar{1}$
				.10697	29.21	012
9 *R*	29.28	3.050	.10749	.10768	29.30	$40\bar{1}$
5 *R*	29.95	2.983	.11235	.11232	29.95	400
8 *R*	30.07	2.972	.11324	.11327	30.07	202
				.11430	30.21	$21\bar{2}$
				.12437	31.55	112
				.13526	32.94	$40\bar{2}$
				.13739	33.21	311
2 *R*	33.38	2.684	.13879	.13903	33.41	$31\bar{2}$
6 *R*	33.76	2.655	.14188	.14195	33.77	$20\bar{3}$
2 *R*	34.15	2.625	.14508	.14499	34.14	003
5 *R*	34.66	2.588	.14931	.14918	34.65	401
				.15021	34.77	$41\bar{1}$
				.15485	35.29	410
2 *R*	35.42	2.532	.15597	.15580	35.40	212
5 *R*	37.07	2.423	.17032	.17012	37.05	020
				.17714	37.83	120
				.17779	37.90	$41\bar{2}$
1 *R*	38.21	2.353	.18056	.17897	38.03	$11\bar{3}$

TABLE 1—(*continued*)

Qual.	2θ obs. Cu K_α	*d* obs.	Q obs.	Q calc.	2θ calc. Cu K_α	*hkl*
2 *R*	38.65	2.327	.18459	.18448	38.64	$21\bar{3}$
				.18623	38.83	021
				.18752	38.97	013
				.18806	39.03	$12\bar{1}$
1 *R*	39.45	2.282	.19200	.19171	39.42	411
3 *R*	39.76	2.265	.19491	.19506	39.78	$40\bar{3}$
				.19820	40.11	220
				.19844	40.13	121
				.20127	40.43	312
				.20393	40.71	$22\bar{1}$
				.20401	40.72	$31\bar{3}$
1	40.78	2.211	.20406	.20419	40.74	203
				.20821	41.15	$51\bar{1}$
1 *R*	41.42	2.178	.21079	.21011	41.35	113
				.21803	42.16	510
5	42.20	2.140	.21844	.21826	42.18	402
				.22469	42.83	221
3	43.38	2.084	.23023	.23059	43.41	$51\bar{2}$
				.23120	43.48	$11\bar{2}$
				.23330	43.68	320
				.23384	43.74	$32\bar{1}$
				.23456	43.81	022
2	44.10	2.052	.23755	.23759	44.10	$41\bar{3}$
				.23770	44.11	$60\bar{1}$
				.24189	44.52	$22\bar{2}$
				.24434	44.76	$20\bar{4}$
				.24672	44.99	213
				.25196	45.49	122
1 *R*	45.55	1.990	.25259	.25272	45.56	600
Fifteen reflections calculated						
				.29744	49.68	$61\bar{2}$
9	49.90	1.826	.29992	.29995	49.90	601
				.30029	49.93	014
				.30122	50.02	$31\bar{4}$
				.30434	50.29	$60\bar{3}$

* Qual. refers to the quality (not the height) of the peak; it estimates, on a scale of 1 to 10, the accuracy with which the peak can be located. *R* stands for "resolved" and indicates that only one reflection can account for the peak.—$Q=1/d^2$ (Q stands for "quadratic"). Q calc. and 2θ calc. are computed from the unit-cell dimensions given at the head of the table.

group, uniquely determined by x-rays, is $P2_1/a$. The cell dimensions, refined by means of a powder pattern also taken by H. Yoder (Table 1), are: $a=12.307\pm0.005$, $b=4.849\pm0.003$, $c=8.124\pm0.004$ Å and $\beta=104.12\pm0.005°$. With 4 formula units per cell, the calculated density is 2.542 g/cm³. The observed density is 2.57 ± 0.10 g/cm³, the high uncertainty being due to the small size of the crystal—0.4 mg.

Our powder data differ sufficiently from those given in the literature (Burkhart and Imhoff, 1947) to warrant careful comparison. Table 2 shows the results of this analysis and explains the discrepancies.

Lithium Disilicate

As reported to us by Kracek $Li_2Si_2O_5$ was crystallized from a melt and annealed at 975° C. It melts incongruently at about 1033° C. to a more siliceous liquid and crystals of $Li_2Si_2O_3$. A rapid high-low transformation takes place at $936\pm3°$; the high-temperature form cannot be quenched. We are therefore dealing with crystals of the low-temperature form.

Many of the "crystals" examined turned out to be glass. The true crystals are small {010} plates, slightly elongated along [001], less than 0.1 mm. long, with striations parallel to [001] and [100]. Merwin (in Kracek, 1930*b*) writes: "*Lithium disilicate* $Li_2Si_2O_5$ appears to be orthorhombic with three cleavages at right angles. One cleavage is micaceous, the other two are nearly perfect. The plane of the optic axes is parallel to the micaceous cleavage, and γ parallel to the intersection of the two best cleavages. Optical character is positive, $2V=50°$ to $60°$. Refractive indices: $\alpha=1.547$, $\beta=1.550$, $\gamma=1.558$, all probably within ±0.001."

Weissenberg and precession photographs yield the diffraction aspect $Ccc*$, compatible with space groups $Ccc2$ and $Cccm$. The cell dimensions are: $a=5.80\pm0.02$, $b=14.66\pm0.05$, $c=4.806\pm0.015$ Å. With 4 formula units per cell, the calculated density is 2.438 g/cm³, as compared with an observed density of 2.454 g/cm³ at 25° C. (Jaeger and van Klooster, 1914).

Lithium Metasilicate

Only one form of Li_2SiO_3 exists, with melting point $1201\pm1°$ C. (Kracek, 1930*b*). The compound was correctly described as orthorhombic pseudo-hexagonal by Hautefeuille and Margottet (1881), yet the Barker Index (1952) still lists it as hexagonal.

Li_2SiO_3 was melted, crystallized, and annealed at 1115° (Kracek). Most of the crystals are lath-shaped, less than 1 mm. in length, tabular on {010} and elongated along [001]; their faces give poor reflections. A few crystals, however, short-prismatic in habit, could be measured on the goniometer and proved to be pseudo-hexagonal prisms.

TABLE 2. ANALYSIS OF POWDER DATA IN LITERATURE IN THE LIGHT OF PRESENT RESULTS

hkl	$d_{\text{Lit.}}$*	$I_{\text{Lit.}}$*	*d* calc.	Peak hght. obs.	Remarks
—	6.61	vw	—	—	FeK$_\beta$, *d*(200) = 6.00
200	5.98	ms(4)	5.967	77	Misprint (not 3.98)
$20\bar{1}$	5.46	vw	5.439	6	
—	4.72	vw	—	—	FeK$_\beta$, d(201) = 4.28
110	4.57	vw	4.492	—	
201	4.26	m	4.281	25	
011	4.13	w	4.129	7	
002	3.94	m	3.940	42	
210	3.76	w	3.764	6	
111			3.757		
$21\bar{1}$	3.63	ms(6)	3.619	18	
—	3.52	vw	—	—	Unaccounted for.
—	3.36	w	—	—	FeK$_\beta$, *d*($11\bar{2}$, $40\bar{1}$) = 3.06
$11\bar{2}$	3.07	m(5)	3.107	3	
310			3.076	—	
$31\bar{1}$			3.068	—	
012			3.058	—	
$40\bar{1}$			3.047	41	
400	2.96	s(2)	2.984	47	
202			2.971	58	
112	2.85	w	2.836	—	
$31\bar{2}$	2.67	m(7)	2.682	5	
$20\bar{3}$			2.654	23	
401	2.58	m(8)	2.589	21	
$41\bar{1}$			2.580	—	
212	2.52	w	2.533	5	
020	2.42	m(9)	2.425	10	
$11\bar{3}$	2.35	vw	2.364	3	
$21\bar{3}$			2.328	3	
411	2.27	w	2.284	3	
$40\bar{3}$			2.264	6	
$51\bar{1}$	2.19	vw	2.192	—	
113			2.182	2	
510	2.14	m(10)	2.141	—	
402			2.140	13	
$51\bar{2}$	2.08	w	2.082	4	
$41\bar{3}$	2.05	w	2.052	4	
600	1.99	w	1.989	2	
Sixteen reflections calculated, four of them observed by B. and I.					
601	1.82	s(1)	1.826	28	
etc. . . .					

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* $d_{\text{Lit.}}$ and $L_{\text{Lit.}}$ refer to interplanar distances and intensities given by Burkardt and Imhoff (1947), who used unfiltered iron radiation. $d_{\text{calc.}}$ is computed from $Q_{\text{calc.}}$ of Table 1. "Peak height observed" refers to height above background on the chart, expressed on an arbitrary scale.

Weissenberg and precession photographs show that the crystal system is orthorhombic with diffraction aspect $Cc{*}{*}$, which permits the space groups $Ccmm$, $Cc2m$ or $Ccm2$. The cell dimensions are $a=5.43\pm0.02$, $b=9.41\pm0.03$, $c=4.660\pm0.015$ Å. It follows that $a/b=0.577\pm0.004$ is equal to $\tan 30°=0.5774$ and that the lattice, from a purely metric point of view, is hexagonal, within the experimental accuracy. With 4 formula units per cell, the calculated density is 2.50 g/cm^3, in agreement with the observed density of 2.520 g/cm^3 at 25° C. (Jaeger and van Klooster, 1914).

The lath-shaped irregular crystals show parallel extinction and positive elongation (on the sign of the elongation, we confirm Jaeger and van Klooster, 1914, and disagree with Merwin, in Kracek, 1930*b*). In convergent light a flash figure[2] is obtained, which yields the following optical orientation: $\gamma=c$, $\alpha=a$, $\beta=b$. The crystals are biaxial positive, with medium birefringence. Judging from the pseudo-uniaxial character (Merwin, *op. cit.*), 2V must be small. The refractive indices on record are as follows: $n_\gamma=1.609\pm0.004$, $n_\alpha=1.584\pm0.002$, $B=0.025$ (Jaeger & van Klooster, 1914); $n_\epsilon=n_\gamma=1.611\pm0.001$, $n_\omega=n_\alpha$ or $n_\beta=1.591\pm0.001$ (Merwin, *op. cit.*)

The crystal structure of Na_2SiO_3, the corresponding sodium salt, has recently been described by Grund and Pizy (1952), who kindly let us see their manuscript prior to publication. It consists of single chains of SiO_4 tetrahedra, parallel to the c axis and held together by sodium ions. The structure differs from that of the pyroxenes in that it has four alkali ions per Si_2O_6 link of the chain rather than two alkaline earth ions. Grund and Pizy remark that each sodium ion is coordinated to only five oxygens at distances ranging from 2.27 to 2.45 Å. Note, however, that one of these oxygens is a "dummy," as it is shared between two silicons and is therefore electrostatically neutralized. The coordination number is actually four, amazingly low for sodium. With four sodium-oxygen bonds of strength $\frac{1}{4}$, the Pauling rule is perfectly satisfied since each unshared oxygen receives four such bonds. The coordination polyhedron of sodium roughly approximates a tetrahedron.

Since the lithium ion is smaller than the sodium ion, Li_2SiO_3 and Na_2SiO_3 may be expected to have isotypic structures. Indeed the space group reported for the sodium compound is (in our orientation, with $a<b$) $Ccm2$. Its cell dimensions, $a=6.02$, $b=10.43$, $c=4.81$, are larger than those of the lithium silicate, a and b by 10.9% and 10.8% respectively, c by only 3.2%. The ratio $a/b=0.5772$ is also that required by a

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[2] The isogyres flash out in less than 16° rotation. The possibility of an obtuse bisectrix figure is not absolutely ruled out; in this alternate interpretation, $\alpha=b$, $\beta=a$.

hexagonal lattice. The increases in the *a* and *b* lengths as one goes from Li to Na are in keeping with the increase in ionic diameter; the *c* length changes less as it is mostly governed by the SiO_3 chains. The structural hypothesis of chains stretching along the *c* axis is also supported by the optical character of Li_2SiO_3, biaxial positive with small 2V and acute bisectrix parallel to *c*.

In the system Na_2SiO_3–Li_2SiO_3 Kracek (1939) finds an intermediate compound $NaLiSiO_3$, complete solid solution between it and Na_2SiO_3, and limited solid solution between it and Li_2SiO_3. The crystal structures of Li_2SiO_3 and $NaLiSiO_3$ have not as yet been determined.

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OBSERVATIONS ON THE DISTRIBUTION OF TRACE ELEMENTS IN THE PERTHITE PEGMATITES OF THE BLACK HILLS, SOUTH DAKOTA

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ABSTRACT

The trace-element contents of the Black Hills perthites have been determined spectrographically. Their chief trace elements are found to be Rb, Ba and Sr. The distribution and relative proportions of the different trace elements are discussed. They seem to support the metasomatic derivation of the perthites from the schists of the country rocks in which the pegmatites occur.

STATEMENT OF PROBLEM

The perthite pegmatites of the Black Hills, South Dakota, have been considered by the writer (Higazy, 1949) to have been formed not by magmatic crystallization but mainly by replacement processes. The chemical composition of these perthites as well as their textural and microscopic features support their metasomatic derivation. Moreover, field observations seem also to be in favor of the replacement origin of these pegmatites. The trace-element contents of the perthites and of some of the country rocks were determined spectrographically in order to study the significance of such elements in the petrogenesis of these perthites.

SPECTROGRAPHIC ANALYSIS AND DATA OBTAINED

Portions of the powders used previously for chemical analysis were spectrographically analyzed by the writer at the Macaulay Institute for Soil Research at Aberdeen, Scotland, using the semiquantitative method described in detail by Mitchell (1948). The determined trace elements, and the wave lengths of their respective diagnostic lines have already been recorded by the writer (Higazy, 1952*a*).

The trace-element contents of the analyzed perthites and cleavelandite are listed in Table 1; and those of the other country rocks are given in Table 2. For the exact location of these rocks, and their chemical composition and petrographic description, reference should be made to an earlier paper (Higazy, 1949, Fig. 1, p. 556, p. 559 and pp. 562–564).

MAJOR ELEMENTS

The average ratio Or:Ab:An in the analyzed perthites is 75.8:22.4:1.8 (Higazy, 1949, p. 561). The most important major elements in their composition are, therefore, Si, Al and K. Sodium and Ca may also be significant but not to the extent of that of K. The ranges of K_2O, Na_2O and CaO in these perthites are 10.58–15.62, 0.38–3.29 and 0.15–0.82 per cent, respectively.

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Small amounts of Mg and Fe are usually present in the composition of perthites. MgO and total iron oxides (as Fe_2O_3) in the perthites of the Black Hills vary from 0.00–0.12 and 0.06–0.13 per cent, respectively. The exact mode of occurrence of these elements, namely, Mg, Fe^2 and Fe^3, in the structure of potash feldspar is not yet perfectly known. Among the major elements of the perthite, Na possesses the closest ionic radius (0.98 Å) to that of Mg (0.78) and that of Fe^2 (0.83). It is not certain that Mg and/or Fe^2 replace Na in the feldspar lattice since (*a*) Na usually has a higher coordination number and (*b*) Mg and Fe^2 appear to have much higher activation energy values (E-value). Magnesium is known to replace Al in biotite and hornblende. Although these species have different structures from that of the feldspars, yet the possibility of such replacement in the latter should not be disregarded. If, however, Mg and Fe^2 have no definite position in the structure of the feldspar, it may then be conjectured that they are either dispersed through the structure or present in minute mineral inclusions in the perthite. A combination of these two alternatives is also a possibility.

Fe^3 has an ionic radius of 0.67 Å, not very different from that of Al^3 (0.57), and both have the same valency. If the feldspar lattice is favorable for the admission of Fe^3, it may be surmised that it replaces Al, since Wickman (1943, p. 381) points out that Fe^3 may be considered as so immobile that it could be incorporated in a suitable lattice under any conditions.

TRACE ELEMENTS

The different trace elements in the perthite pegmatites are classified according to their abundance into four categories: abundant, present in small amounts, rare and very rare.

ABUNDANT: Rb, Ba and Sr.
PRESENT IN SMALL AMOUNTS: Li, Pb, Ga and Tl.
RARE: Mn, V, Cu; Be, Ge and Zr.
VERY RARE: Cr, Co, Ni, Sc; Sn, Ag, Mo; La, Y and In.

Rubidium, barium and strontium

The Rb content in the analyzed perthites (Nos. 1–15) varies from 600 to 9500 ppm. and the average is 2665 ppm. (Table 1). The average in the two analyzed microcline-perthites from the Hugo pegmatite is 3600 ppm. The Rb content of the potash feldspar of the Caledonian granodioritic rocks of Scotland studied by Nockolds and Mitchell (1948); Table XXI, p. 571) varies from 60 to 800 ppm., the highest being in potash feldspar from aplite, Garabal Hill-Glen Fyne Complex, Scotland.

TABLE 1. TRACE-ELEMENT CONTENTS IN PPM. OF THE ANALYZED PERTHITES AND CLEAVELANDITE

Element	Sensitivity	1	2	3	4	5	6	7	8	9	10	11
Rb	1	2800	1500	2500	5000	1000	600	800	5000	9500	6000	800
Li	1	40	4	9	7	25	3	2	170	300	100	8
Ba	5	45	80	25	260	200	400	1%	1000	20	830	1500
Sr	5	1500	12	20	180	45	70	750	400	600	90	60
Cr	1	*	*	*	*	*	*	*	*	5	3	3
Co	2	*	*	*	*	*	*	*	*	*	*	*
Ni	2	*	*	*	*	*	*	*	*	*	*	*
Zr	10	*	*	*	*	300	*	*	*	*	*	*
La	30	*	*	*	*	*	*	*	*	*	*	*
Y	30	*	*	*	*	*	*	*	*	*	*	*
Cu	3	3	3	4	5	4	3	3	3	4	4	3
V	5	5	5	*	5	*	5	5	5	20	10	10
Ga	1	35	30	40	35	50	30	30	40	45	30	30
Tl	30	30	*	30	65	40	*	*	140	150	80	*
Sn	5	*	*	*	*	*	*	*	*	*	*	*
Pb	10	10	18	15	50	15	30	35	20	40	10	40
Sc	10	*	*	*	*	*	*	*	*	*	*	*
Mn	5	8	10	10	10	10	12	8	7	8	10	10
Mo	1	*	*	*	*	*	*	*	*	1	*	*
Ge	10	*	*	*	*	*	*	*	15	10	*	*
Be	5	8	*	*	10	*	*	*	*	8	*	8
Ag	1	*	*	*	*	1	*	*	*	1	1	3
In	10	*	*	*	*	*	*	*	*	*	*	*

Nos. 1 to 11 correspond to numbers of chemical analyses by R. A. Higazy recorded by the writer (Higazy, 1949, p. 559).

Spectrographic analyst: R. A. Higazy.

* Element if present is in amounts below its limit of sensitivity.

But the perthite found as phenocrysts in the granites of the Glendale area of the Black Hills (analysis 13, Table 1) has 900 ppm. Rb. On the whole, the Rb content of the Black Hills perthites seems to be higher than that of the Caledonian feldspars.

Ahrens (1945*b*) examined the Rb content of 67 specimens of potash feldspars (microcline, orthoclase and perthite). Most of these, however, were collected from pegmatites of widely different occurrence. Their Rb content varies from 40 to 2560 ppm., with an average of 295 ppm. The range in the 16 green microcline (amazonite) examined also by him, is from 640 to 11,880 ppm., with an average of 3670 ppm. The average in perthites of the Black Hills (2665 ppm.) is not appreciably lower than that of the green microcline. Among the fifteen analyzed perthites of

Table 1. Trace-Element Contents in ppm. of the Analyzed Perthites and Cleavelandite—(*Continued*)

Element	Sensitivity	12	13	14	15	Average	16	17	Cleavelandite
Rb	1	2000	900	1000	600	2665	1200	6000	2
Li	1	130	25	55	2	60	35	160	150
Ba	5	100	150	200	100	965	30	550	5
Sr	5	240	25	50	35	270	800	90	100
Cr	1	3	1	3	1	1.5	1	1	1
Co	2	*	*	*	*	*	*	*	*
Ni	2	*	*	*	*	*	*	*	*
Zr	10	*	*	*	*	*	*	*	30
La	30	*	*	*	*	*	*	*	*
Y	30	*	*	*	*	*	*	*	*
Cu	3	3	5	4	3	4	4	4	5
V	5	5	5	8	5	6	5	15	10
Ga	1	30	30	35	30	35	35	40	70
Tl	30	45	*	*	*	39	*	80	*
Sn	5	*	*	*	*	*	200	*	*
Pb	10	20	25	20	450	53	10	10	*
Sc	10	*	*	*	*	*	*	*	*
Mn	5	15	10	12	10	10	12	8	40
Mo	1	*	*	*	*	*	*	1	*
Ge	10	10	*	*	*	*	*	10	20
Be	5	*	*	*	*	*	8	*	120
Ag	1	*	1	4	1	1	3	1	*
In	10	*	*	*	*	*	*	*	*

Nos. 12 to 17 correspond to numbers of chemical analyses by R. A. Higazy recorded by the writer (Higazy, 1949, p. 559).

Spectrographic analyst: R. A. Higazy.

* Element if present is in amounts below its limit of sensitivity.

the Black Hills, one is grayish in color (No. 9), five are grayish-white (Nos. 2, 4, 8, 10 and 12) and the other nine are pinkish (Higazy, 1949, p. 563). The grayish perthite is relatively the richest in Rb (9500 ppm.). The average Rb content of the six grayish and grayish-white samples is 4835 ppm., while that of the nine pinkish perthites is only 1220 ppm. This seems to be a fairly reliable indication that the pink varieties of potash feldspar contain relatively less Rb than the grayish types. The average Rb of the grayish perthites (4835 ppm.) lies within the range given by Ahrens (1945 *b*, p. 229) for amazonite (4570–5485 ppm.) and closer to the minimum limit, while the average for the pink types is notably less than for the green species being, only 1220 ppm.

Barium in the Black Hills perthites is less abundant than Rb except

TABLE 2. CHEMICAL COMPOSITION AND TRACE-ELEMENT CONTENTS OF THE COUNTRY ROCKS

	A	B	C	D
	Biotite-schist	Perthite-schist	Granite	Albitite
Chemical composition as weight per cent of the oxide				
Oxide				
SiO_2	68.83	70.85	73.48	69.93
TiO_2	0.46	0.21	0.10	0.00
Al_2O_3	14.63	15.48	15.10	17.84
Fe_2O_3	1.42	0.64	0.41	0.14
FeO	3.83	1.09	0.82	0.20
MnO	0.05	0.01	0.02	0.01
MgO	2.24	0.83	0.60	0.54
CaO	0.52	0.38	0.80	0.59
Na_2O	1.05	2.16	3.34	9.68
K_2O	5.01	6.45	4.31	0.42
P_2O_5	0.17	0.45	0.13	0.51
H_2O	1.52	1.18	0.94	0.23
H_2O^-	0.14	0.08	0.08	0.06
Trace elements as parts per million of the element				
Element				
Rb	240	600	240	1
Li	30	70	150	10
Ba	2000	3500	750	25
Sr	35	150	50	5
Cr	80	60	3	4
Co	10	3	*	2
Ni	40	20	*	*
Zr	220	250	250	50
La	30	30	30	*
Y	120	60	40	30
Cu	9	3	8	6
V	50	60	20	12
Ga	35	40	50	60
Tl	*	*	*	*
Sn	5?	*	5?	*
Pb	12	18	12	*
Sc	*?	*	*	*
Mo	1	*	*	*
Ge	*?	*	*	*
Be	5	8	25	10
Ag	<1	<1	<1	<1
In	*	*	*	*

Chemical and Spectrographic analyst: R. A. Higazy.

Chemical analyses A, B and D are taken from Higazy (1949, p. 560).

* Element if present is in amounts below its limit of sensitivity given in Table 1.

in the perthite porphyroblast (No. 7) found in the perthitized schist. It is interesting to notice that the color of this Ba-rich perthite is salmon red and that it contains 0.13 per cent of total iron oxides as Fe_2O_3, an amount similar to that of perthite No. 15, which is pinkish. Moreover, perthites Nos. 10 and 11 are grayish-white and deep pink, respectively; yet each has 0.09 per cent Fe_2O_3. But the Ba content of the grayish-white perthite (380 ppm.) is less than that of the deep pink (1500 ppm.). The reddish coloration, therefore, seems to be related to relative enrichment in Ba rather than in Fe.

The Ba content ranges from 20 to 10,000 ppm. The highest abundance for the pegmatitic perthites excluding the perthite porphyroblast of the schist is 1500 ppm., while the average is 965 ppm. The average Ba content of the microcline-perthites of the Hugo pegmatite is 290 ppm. In the potash feldspars of the Caledonian granodioritic rocks Ba ranges from 2000 to 8000 ppm. and is more abundant than Rb, as in the porphyroblast of the perthitized schist of the Glendale area of the Black Hills. But the Ba content (50 ppm.) of the potash feldspar from aplite, Garabal Hill-Glen Fyne Complex, (Nockolds and Mitchell, 1948, p. 571) is lower than that of Rb (2000 ppm.), as in the pegmatitic perthites and the perthite phenocryst of the granite of the Black Hills. Although the data for the Black Hills perthites do not show a regular relationship between Rb and Ba, yet they suggest that the relatively Rb-rich perthite (No. 9) possesses a relatively low content of Ba.

Strontium varies from 20 to 1500 ppm., with an average of 270 ppm. It is less abundant than Ba in most of the analyzed perthites, as in the potash feldspars of the Caledonian rocks and those of the pre-Cambrian granites of Boulder Creek, Overland Mtn. and Silver Plume, Jamestown, Colorado, examined by Bray (1942).

That Rb, Ba and Sr are concentrated in the K-rich feldspars is very obvious, especially when comparison is made with the content of these three elements in cleavelandite. This Na-rich species has only 0.09 per cent K_2O and its Rb, Ba and Sr contents are 2, 5 and 100 ppm., respectively. Rubidium and Ba are thus present in negligible amounts compared with their respective average abundances in the perthites. Strontium however, is relatively high. But cleavelandite has 0.50 per cent CaO. Its Or:Ab:An ratio is 0.6:97.0:2.4 (Higazy, 1949, p. 561). It is quite likely that Sr (1.27 Å) substitutes for Ca (1.06) in cleavelandite.

Lithium, lead and gallium

Lithium is present in all the analyzed perthites but in widely variable amounts, the range being from 2 to 300 ppm. and the average 60 ppm. It seems that the relatively Rb-rich perthites (Nos. 8, 9 and 10 of Glendale area pegmatites and No. 16 of the Hugo pegmatite) are also rich in Li. But this relationship cannot be generalized since perthite No. 12 has 2000 ppm. Rb and 130 ppm. Li, while perthite No. 4 has 5000 ppm. Rb and only 7 ppm. Li. Moreover, the potash feldspar of the adamellite, Morven-Strontian Complex, Scotland (Nockolds and Mitchell, 1948), has 600 ppm. Rb and 8 ppm. Li, while the range of the Rb content of the other Caledonian potash feldspars is from 60 to 2000 ppm., and their Li is below its limit of sensitivity (1 ppm.).

The position of Li in the feldspar lattice is debatable. Lithium feldspars do not exist in nature, and according to Ahrens (1945 *a*, p. 82), a Li content greater than 2–5 ppm. is exceptional in the feldspars. However, the Li content of the micas is generally appreciable. There are good grounds for assuming that Li replaces Mg in biotite. Both elements have the same ionic radius (0.78 Å) and the same coordination number. On the other hand, muscovite does not usually contain Mg. But Hendricks (1939, p. 763) mentions that Li can enter the muscovite structure giving rise to lithium muscovite. Moreover, Ahrens finds as much as 1.2 per cent Li_2O in the muscovites of Uranoop River Area, Namaqualand, S. Africa (Ahrens, 1945*a*). He suggests that Al may be replaced by Li in the muscovite structure providing that electrostatic balance is restored by a concomitant replacement elsewhere in the lattice. The proposed substitution of Al by Li in muscovite may possibly imply that such replacement could also take place in the feldspar lattice, as Lundegårdh suggests (1947, p. 108). It should also be mentioned that Li (0.78 Å) may replace Na (0.98). Cleavelandite has 150 ppm. Li, but this Na-rich species also possesses a high amount of Al. On the other hand the analyzed albitite (Table 2) is relatively poor in Li (10 ppm.). This makes the replacement of Na by Li rather uncertain.

Lead is present in all the analyzed perthites. Omitting No. 15 it ranges from 10 to 40 ppm., with an average of 25 ppm. Perthite No. 15 is exceptionally high in Pb (450 ppm.) and if it is taken into account, the average rises to 53 ppm. It is interesting to notice that the total of Rb, Ba and Sr in the exceptionally high Pb perthite (No. 15) is the lowest among all the samples tested. Pb ores were not detected microscopically in the thin section of this perthite. Lead is below its limit of sensitivity (10 ppm.) in cleavelandite. It is most likely replacing K in the lattice of the potash feldspars; the association of Pb and K in silicic rocks has been pointed out by Sandell and Goldich (1943, p. 169).

Gallium occurs in the analyzed samples rather uniformly, the range being from 30 to 50 ppm. and the average is 33 ppm. Cleavelandite has 70 ppm. Ga. The geochemical relationship between Ga and Al is unique.

Gallium feldspars, $KGaSi_3O_8$ and $NaGaSi_3O_8$, in which Ga completely replaces Al, have been synthesized by Goldsmith (1950).

Thallium

The abundance of Tl in the perthites of the Black Hills is very interesting. The average Tl content of the fifteen perthites of the Glendale area is 39 ppm. It is found in amounts below its limit of sensitivity (30 ppm.) in some perthites and in others it may be present up to 150 ppm. Shaw (1952, p. 128) reports 84 and 21 ppm. Tl in the microcline-perthites of Custer Mt. and the High Climb pegmatites of the Black Hills, respectively. The microcline of the Hugo pegmatite has 80 ppm. Tl. In the five perthites which possess less than 1000 ppm. Rb (Nos. 6, 7, 11, 13 and 15) Tl is below its limit of sensitivity. Moreover, the lepidomelane-skarn of Co. Donegal, Ireland, examined by the writer (Higazy, 1952*a*) has as much as 800 ppm. Rb and its Tl content is undetectable. Of the four perthites (Nos. 2, 5, 14 and 16) in which Rb varies from 1000 to 1500 ppm. thallium is present in detectable amount (40 ppm.) only in No. 5. All the remaining perthites (Nos. 1, 3, 4, 8, 9, 10, 12 and 17) contain appreciable amounts of Tl together with at least 2000 ppm. of Rb. This clearly shows that there is a geochemical relationship between Rb and Tl as suggested by Ahrens (1948). It should be stated that Cs (8079.02 Å line) was detected qualitatively in only the spectrum of the perthite No. 9 which possesses the relatively highest contents of both Rb (9500 ppm.) and Tl (150 ppm.). There may be, therefore a relationship between Cs on the one hand and Rb and Tl on the other, especially if we consider that the cesium mineral pollucite examined by Ahrens (1945*b*) also contains appreciable amounts of both Rb and Tl. However, more data from several other Rb-rich potash feldspars are needed to establish the suggested relationship.

The Rb/Tl ratio in the perthites of the Black Hills varies from 25 to 93 with a mean of 60. Ahrens (1945*b*) concluded that the mean weight ratio Rb_2O/Tl_2O is 110 and that the limits of variation are 30 and 650. He also mentions that this ratio can vary in the same type of mineral from one area to another. The mean value of this ratio for the perthites of the Black Hills is 64, with variation from 28 to 98. The log. per cent relationship between Rb_2O and Tl_2O is represented graphically in Fig. 1. The results are practically the same as those plotted by Ahrens (1945*b*, Fig. 3) and the slope of the graph in both cases is nearly 45°.

Rubidium and Tl have the same ionic radii (1.49 Å) and are concentrated in potash-rich minerals, where they enter into the lattices with about the same ease, replacing K.

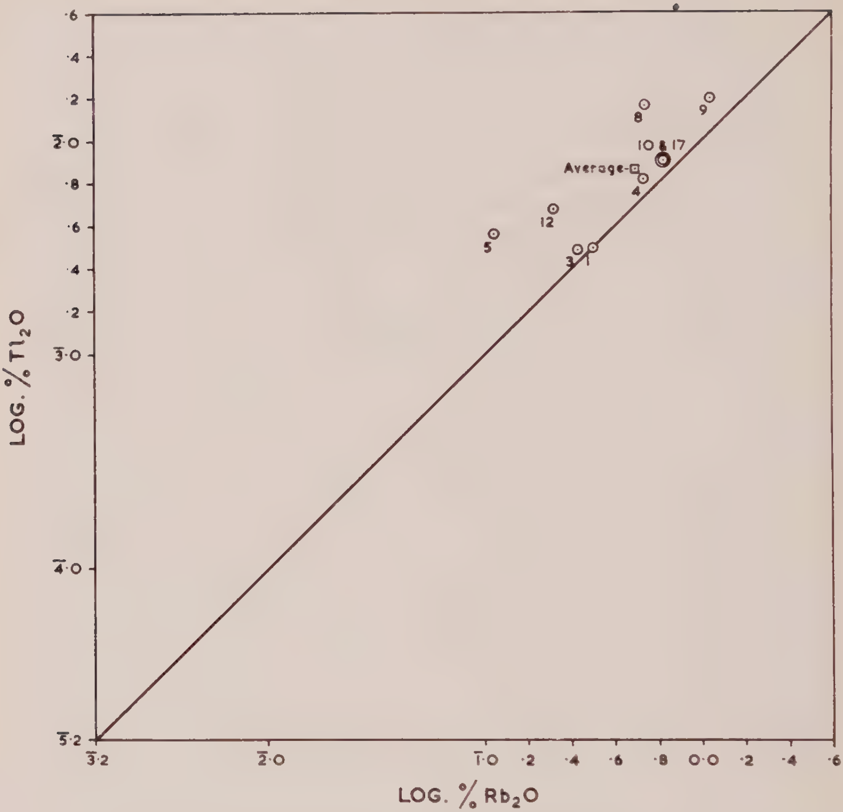

FIG. 1. Relationship between Rb and Tl in the analyzed perthites of the Black Hills.

Manganese, vanadium and copper

Each of the elements Mn, V and Cu is found in almost uniform amounts. Manganese varies from 7 to 15 ppm. with an average of 10 ppm.; V from amounts below its limit of sensitivity (5 ppm.) to 20 ppm., with an average of 6 ppm.; and Cu from 3 to 5 ppm. with an average of 4 ppm. The exact position of each of these elements in the lattice of the potash feldspar is not known. Judging from the ionic radii of Mn^2 and Cu^2, 0.91 and 0.83 Å, respectively, it may be surmised that they replace Na (0.98), and that V^3 (0.65) substitutes for Al (0.57).

The contents of V (10) and Cu (5 ppm.) in cleavelandite are within the range of these elements in the perthites. However, the Mn content in this Na-rich feldspar (40 ppm.) is much higher than in any of the perthites.

Beryllium and germanium

Be is found in all the analyzed perthites except one (No. 6). It varies from 5 to 8 ppm. Germanium, however, is found in only five perthites of the Glendale area and in the microcline of the Hugo pegmatite. Its range is 10 to 15 ppm. It is likely that both Be and Ge replace Si in the feldspar lattice, since the ionic radius of Be (0.34 Å) and that of Ge (0.44) are close to the radius of Si (0.39). Moreover, Ge feldspars, in which Ge completely replaces Si, have been synthesized by Goldsmith (1950). Cleavelandite is richer in both Be and Ge than the perthites, having as much as 120 ppm. of Be and 20 ppm. of Ge.

Zirconium

Zirconium is found only in one perthite (No. 5) and there it is present in rather high amount (330 ppm.). This makes it very likely that minute inclusions of zircon are responsible for the Zr present in No. 5.

The very rare trace elements

Among the very rare trace elements in the analyzed perthites are Cr, Co and Ni. Chromium is found in eight perthites in amounts varying from 1 to 5 ppm., but below its limit of sensitivity in the others. Cobalt and Ni, if present at all, are below their limits of sensitivity (2 ppm.) in all the analyzed samples. In six of the perthites it can be stated that the relative distribution of Cr, Co and Ni is $Cr > Co = Ni$.

Scandium, if present is in amounts below its limit of sensitivity in all the perthites. Chromium, Co, Ni and Sc are elements which generally replace Mg, Fe^2 and Fe^3 in crystal lattices. The relative deficiency of the perthites in these major elements explains the impoverishment of their replaceable trace elements.

Silver is present in most of the analyzed perthites in amounts ranging from <1 to 4 ppm. The ionic radius of Ag is 1.13 Å, very close to that of K. It is possible that the relatively small amount of this trace element replaces K in the lattice.

Tin is found in detectable amount in only one sample (No. 16), collected from the Hugo pegmatite. The high content of Sn (200 ppm.) present in this perthite appears to be accidental, being attributable to very minute inclusions of Sn minerals, possibly cassiterite.

Molybdenum is found in one perthite from the Glendale area and in another from the Hugo pegmatite, the Mo content in each being only 1 ppm.

Lanthanum, Y and In, if present, are each in amounts below the respective limit of sensitivity.

SIGNIFICANCE OF THE DISTRIBUTION OF THE TRACE ELEMENTS

The petrogenetic significance of the distribution of the various determined elements in the perthites can be studied with regard to (*a*) the composition of the different perthites, and (*b*) the trace-element contents of the associated country rocks: namely, the unaltered schist (quartz-muscovite-biotite-schist), the perthitized schist, the granite and the perthite pegmatites.

(*a*) *The composition of the different perthites*

Bray (1942) estimated certain trace elements in the potash feldspars of both Pre-Cambrian granites and their corresponding pegmatites. These rocks were supposed to be of magmatic origin. He (Bray, 1942, p. 797) remarked that the pegmatite minerals usually contain less of almost every minor element than the corresponding minerals from the parental granitic rocks. The statement that pegmatitic minerals should be relatively less enriched in almost every minor element, is too broad a generalization; it is not in complete harmony with Goldschmidt's principles of the replacement of major by trace elements (Goldschmidt, 1937 and 1945), replacement being a process which is mainly governed by the electrostatic charges and the ionic radii of the elements concerned. According to these principles, there should be certain elements which become relatively concentrated in the pegmatite minerals, provided that such minerals also occur in the parental rocks. In such cases the ratio of the content of the trace element to that of its replaceable major element should be higher in the pegmatite mineral. Rubidium is an example of this, since its ionic radius (1.49 Å) is greater than that of K (1.33). The potash feldspar of aplite from the Garabal Hill-Glen Fyne Complex in Scotland has 2000 ppm. Rb, while the average of this element in the potash feldspars of the granodioritic rocks supposed to have crystallized earlier from the magma is only 425 ppm. (Nockolds and Mitchell, 1948). Moreover, the average Rb_2O content of the potash feldspars from granites studied by Ahrens (1948) is 0.0073 per cent, while that of the pegmatitic species is 0.0319 per cent and that of the hydrothermal microclines from pegmatites is as high as 0.433 per cent. Amongst the perthites of the Black Hills Nos. 6 and 15 have the lowest Rb_2O content, 0.066 per cent. The perthite phenocryst (No. 13) from granite, however, has a higher Rb_2O content (0.099 per cent). The Rb/K ratio of the phenocryst (9.4) is also higher than that of the perthites Nos. 6 and 15 (6.1). Moreover, the perthite porphyroblast of the schist has 0.088 per cent Rb_2O and its Rb/K ratio is 7.7. Supposing, for the sake of the argument, that these perthites were derived from strictly magmatic sources, that is by differentiation of a basic magma, then their Rb contents, as well as the values of their Rb/K ratio, would seem to imply that the pegmatite perthites were formed at an earlier stage than the perthite phenocryst from the granite, and that the porphyroblast was developed at an intermediate stage—conclusions that are obviously not feasible.

There are other pegmatitic perthites, especially Nos. 4, 8, 9 and 10 of the Glendale area and No. 17 of the Hugo pegmatite, which have unusually high Rb contents. These perthites cannot be considered as primary products of magmatic crystallization since such products generally have a much lower Rb content. The relatively high Rb and Li contents in these perthites may support the hypothesis that the latter were formed at the hydrothermal stage during which replacement processes become dominant.

There are also certain features in the distribution of the other trace elements which are not in favor of the magmatic mode of genesis. The Ba content of the phenocryst from granite and the value of its Ba/K ($1000\text{Ba/K}=1.6$) are much lower than those of (*a*) most of the pegmatitic perthites (1000Ba/K ranges from 1.9 to 14.7), and (*b*) of the perthite porphyroblast of the schist ($100\text{Ba/K}=97.0$). But according to Goldschmidt's principles, divalent Ba should be more concentrated relative to K in the early formed crystals. Moreover, Engelhardt (1936) has pointed out that Ba is relatively enriched in potash feldspars which form at the earlier stages of crystallization compared with those formed later.

It is important to trace the source of the replacing materials which gave rise to the pegmatitic perthites. They may be the residual hydrothermal solutions which usually remain at the end of the magmatic history. These solutions subsequently replace the magmatic minerals of the pegmatites. But the relative distribution of Cr, Co and Ni leads to the belief that such magmatic hydrothermal solutions cannot be responsible for the development of the Black Hills perthites. It has been found that the relative abundances of these trace elements in the rocks that form during the end stages of magmatic crystallization is $\text{Co}>\text{Ni}>\text{Cr}$ (Ludegardh, 1949, p. 21), or $\text{Ni}>\text{Co}>\text{Cr}$ (Higazy, 1952*b*). Consequently the hydrothermal magmatic solutions as well as the magmatic pegmatites, or even the granites that crystallized immediately before the hydrothermal stage, should have a lower Cr content than that of either Co or Ni. The question now arises, why the Black Hills pegmatites and granite, if of strictly magmatic ancestry, have the relative distribution $\text{Cr}>\text{Ni}\geqq\text{Co}$ which is inconsistent with this mode of origin. Why do we find wide fluctuations in the Rb, Li and Ba contents of the different perthites? It may be argued that the composition of the hydrothermal replacing solutions varies from place to place in the same locality, that is to say, the solutions may have been Rb-rich in one part of the country rocks and Li-rich in another. But then the separation of such differing solutions from the original magmatic liquor calls for explanation. These problems seem to find their solution if we seek the origin of the replacing material in the country rock itself. In the first place, the analyzed schist has the relative distribution Cr > Ni > Co, like that of the associated granite and pegmatites. In the second place, the trace-element contents in the country schists are quite likely to differ from one part to another. Landergren (1945), studying the boron content of sediments, points out that this content varies horizontally as well as vertically in the stratigraphic column from one district to another. Such local differences in the trace-element contents of the country rocks would undoubtedly show up in the corresponding granitization products. This would be so quite apart from the fact that the mobilities of the different elements also vary and would therefore tend to give rise to different concentrations of such elements in the various parts of the rocks through which they pass. Moreover, it seems that each element would tend to follow preferential paths determined by the structural features of the area and therefore, for this additional reason, the relative concentrations would vary in different parts of the same locality.

(*b*) *The trace-element contents of the various country rocks*

The chemical composition and the trace-element contents of the country rocks, namely, the unaltered schist, the perthite-schist, the granite and the albitite are recorded in Table 2.

It is important to notice that the schist does not lack any of the trace elements present in the granitic country rocks. Moreover, they all have Cr > Ni > Co and Ba > Sr. This is the same relationship as that found in the psammo-pelitic schists of the different metamorphic zones of the Scottish Highlands (Higazy, 1952*c*). Here, also the relative distributions Cr > Ni > Co and Ba > Sr persist in the analyzed rocks of the chlorite-, biotite-, garnet-, staurolite- and sillimanite-zones.

The essential mineral constituents of granites, wherever they occur, are substantially the same. Yet the distribution of the rarer elements in these granites varies rather markedly. The trace-element contents of the Pre-Cambrian granites of the Black Hills (S. Dakota), the rapakivi granite of Finland and the Caledonian granites (adamellites) of Moy and Morven-Strontian (Scotland) are listed in Table 3. It will be noticed that the rapakivi granite has 1000 ppm. Rb, while the Scottish rocks have 500 and 400 ppm., and the Black Hills granite only 240 ppm. Furthermore, both the rapakivi and the Black Hills granites are relatively poor in Sr, having 39 and 50 ppm., respectively, while the Scottish granites have 1200 and 1500 ppm. of this element. The values of Ba/Sr in the granites of Finland (rapakivi), Black Hills, Moy and Morven-Strontian are 12, 15, 2.5 and 1, respectively. These marked differences cannot be entirely attributed to the different ages of these granites, since Sahama and Rankama (1939, p. 8) point out that the Pre-Cambrian granites in the different parts of Finland show a remarkable variation in their rarer-element contents, a discovery which led them to the belief that not all the granites could be of magmatic origin.

TABLE 3. TRACE-ELEMENT CONTENTS IN PPM. OF GRANITES

Element	Rapakivi granite[1]	Black Hills granite	Morven-Strontian, Scotland adamellite[2]	Moy, Scotland adamellite[2]	Average granite[3]
Ba	480	570	1500	3000	445
Sr	39	50	1500	1200	85
Rb	1000	240	400	500	550
Li	25	150	40	25	185
Co	0	<1	<1	10	8
Ni	0	<1	8	10	2
Cr	28	3	10	20	2
Ga	75	50	15	15	15
V	2	20	70	20	—

[1] Amounts taken from Sahama (1945, p. 59.)

[2] Amounts taken from Nockolds and Mitchell (1948, p. 549 and p. 550).

[3] Amounts taken from Wager and Mitchell (1943, p. 237).

— No data.

It is difficult to see how residual granitic melts which might form at the late stages of differential crystallization of basic magma could exhibit such wide variations in their trace-element contents from one geographical occurrence to another. It is a more reasonable inference from the data available that the granites concerned formed either by metasomatic processes within the country rocks or by a further extension of such processes—leading to refusion, whereby granitic ichors might be formed. Moreover, if the normative composition of several granitic and pegmatitic rocks supposed to have crystallized at the late stages of the differentiation history of basic magma be plotted on the orthoclase-albite-anorthite and the $NaAlSiO_4$–$KAlSiO_4$–SiO_2 equilibrium diagrams, the distribution is inconsistent with the hypothesis of their derivation from a differentiating basic magma (Higazy, 1950). The normative feldspar contents lie in the orthoclase field of the orthoclase-albite-anorthite equilibrium diagram and their salic normative constituents place them outside the region of low temperature in the $NaAlSiO_4$–$KAlSiO_4$–SiO_2 equilibrium diagram. It has been generally assumed that the discrepancy between the chemical composition of such rocks and the results of high temperature studies of analogous systems is mainly due to the effect of water vapor which usually concentrates towards the end of the magmatic history. However, this explanation is considered, on theoretical grounds, to be inadequate (Higazy, 1949). Moreover, Bowen and Tuttle (1950, p. 497) have recently established that water vapor has no discernible effect on the fundamental course of crystallization in the albite-orthoclase binary system i.e., in the system which is the most important one concerned in the problem of the perthite.

The discrepancy between the normative composition of the Black Hills perthites and the orthoclase-albite-anorthite equilibrium diagram, together with the textural features, were considered as criteria supporting a metasomatic origin for these perthites (Higazy, 1949). Metamorphic differentiation, which is responsible for the development of the perthite porphyroblasts in the schist, was thought to give rise to chemically similar perthite pegmatites in parts of the country rock where favorable physical and structural conditions prevailed. Experimental data, given by Jagitsch, (1949) however, suggested that the depth affected by solid diffusion is only a few metres in 10^8 years. Although so slow a rate could satisfactorily account for the formation of the perthite porphyroblasts yet it cannot explain the development of large pegmatitic bodies by "solid" diffusion. But the state under which diffusion takes place may not necessarily be that of a perfect solid; moreover, we do not yet know the exact effect on diffusion rates of the very complicated pressures and stresses that attend orogenesis. Such conditions may change an apparently solid rock into a state concerning which our knowledge is so far negligible (Bridgman, 1951). However, it is quite common to find appreciable amounts of water in sediments and especially in the pelitic types. Garrels, Dreyer and Howland (1949) report a rate of penetration for ions during their diffusion through intergranular spaces in water-saturated rocks, of two and a half miles in a million years at a temperature of 100° C. Moreover, they mention that the rate of diffusion increases with temperature. These findings add strength to the view that water might have accelerated the reactions and transformations during the metamorphic differentiation of the country schist which is regarded as having given rise to the perthites (Higazy, 1949, p. 580).

The significance of isotopic composition of the oxygen in silicate rocks in petrogenesis has been pointed out by Beartschi (1950). Silverman (1952) has recently determined the relative abundances of the oxygen isotopes in different types of rocks. He has found that granites have higher O^{18} abundances compared to the basic rocks. This fact implies that the granites he investigated are not late crystallizing fractions of a differentiating basic magma. Furthermore, he finds that the rocks which are supposed to be the late differentiates have variable isotopic abundances which can be correlated with variation in their parental sedimentary materials (Silverman, 1952, p. 35). The variation in the trace-element contents of the perthites of the Black Hills and the appreciable differences of these contents in the granites of various geographical regions can also be ascribed to corresponding variations in the lithological and chemical constitution of the original country rocks already noted.

Gains and Losses of the Trace Elements During the Metasomatism of the Country Schist

If the hypothesis of the metasomatic derivation of the Black Hills perthites is true, it is of special interest to ascertain the behavior of the trace elements during the reorganization of the atoms and molecules inside the schist to give rise to these perthites. The additions and subtractions of the major elements which were inferred to have taken place during the metasomatism have already been given by the author (Higazy, 1949, Table 8, p. 574).

The metasomatic alteration of the quartz-muscovite-biotite-schist to perthitic schist comprised:

addition of K, Na, Al and Si,
and subtraction of Mg, Fe^2, Fe^3 and Ca.

Reference to Table 2 shows that, as regards the trace elements there was

addition of Rb, Ba, Sr and Pb; Ga; Be; Li and Zr,
and subtraction of Cr, Co, Ni, Cu and Mo; and Y.

Thus Rb, Ba, Sr and Pb behaved similarly to K, while Ga followed Al and Be most probably accompanied Si. Lithium usually follows Mg, but in this particular case Li was added while Mg was subtracted. It has already been mentioned that Li may replace Al in the feldspar lattice. Since we here find that both Li and Al were added, it appears that replacement of Al by Li is highly probable. Chromium, Co, Ni, Cu and Mo followed their replaceable major elements Mg, Fe^2 and Fe^3; and Y behaved similarly to Ca.

The geochemical changes of the major elements involved in the transformation of the country schist into average perthite are mainly the same as those concerned in the formation of the perthite-schist. Potassium, Na and Al were added; and Ca, Mg and Fe were subtracted. The K-replacing trace elements which were added are Rb, Sr, Pb and Tl. Again Tl shows its geochemical coherence to Rb. Barium, however, was subtracted; but since Ba can substitute for both K and Ca it seems that it followed Ca in this case, particularly as the gain in K (8.83)* is far greater than the loss in Ca (0.13).* Similarly the total amount of added Rb, Sr, Pb and Tl is greater than the amount of the subtracted Ba. Chromium, Co, Ni, Cu, V and Mo followed Mg, Fe^2 and Fe^3 in being subtracted; and La and Y decreased with Ca.

During the formation of the Black Hills perthites the trace elements therefore behaved as they have been found to do (*a*) in the metasomatic-metamorphic development of skarn rocks from an epidiorite in the Malin Head district of Co. Donegal, Ireland (Holmes and Reynolds, 1947; Higazy, 1952*a*), and (*b*) in the development of the various metamorphic zones of the Scottish Highlands (Higazy, 1952*c*). It is important to mention that the diffusion coefficients (*D*, in cm^2 day) recorded by Garrels, Dreyer and Howland (1949, p. 1819) show that the trace elements have the same or very similar diffusion coefficients as those of their replaceable major elements. Considerable confidence can be placed, therefore, in suggesting the addition of diffusion coefficient as a new significant factor in controlling the distribution and behavior of the trace elements during the progress of metasomatic and metamorphic processes.

CONCLUSIONS

The Black Hills perthites contain appreciable amounts of Rb, Ba and Sr. These elements have approximately the same ionic radii as that of K and are known to substitute for it in the crystal lattices. Thallium is present in detectable amounts in the perthites which are relatively rich in Rb. Lithium, Pb and Ga are present in amounts notably less than those of Rb, Ba and Sr. Manganese, V, Cu, Be and Ge are relatively rare; and Ag, Cr, Co, Ni, Sc, Mo, La and Y are very rare. The pink and reddish perthites seem to be relatively poor in Rb but relatively rich in Ba, while the grayish types show the reverse relationship.

The trace-element contents are variable in the different perthites. The Black Hills granite and perthite pegmatities have $Cr > Ni > Co$ and $Cr > Ni = Co$, respectively, which is not consistent with either a magmatic or a hydrothermal mode of origin. The replacing materials seem to have been derived endogenetically from (within) the country schist which also has $Cr > Ni > Co$.

During the metasomatic development of the Black Hills perthites the trace elements behave similarly to the major elements which respectively have approximately the same diffusion coefficients, and ionic radii and are known to be replaced by them in favorable crystal lattices.

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* Amount of gain or loss is calculated from Barth's (1948) rock formulae (based on a unit cell for each rock of 160 oxygen atoms) and recorded by Higazy (1949, p. 574).

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CHINOITE, A NEW MINERAL

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ABSTRACT

Crystallographic studies of small, emerald-green crystals from the Chino pit, Chino Mines Division, Kennecott Copper Corporation, Santa Rita, New Mexico led to the description of a new mineral dimorphous with pseudomalachite.

X-ray studies by the Weissenberg method on the new mineral showed it to be orthorhombic and gave a space group $Pnn—C_{2v}^{10}$ with cell dimensions $a_0=7.47$ Å, $b_0=8.31$ Å, $c_0=5.83$ Å. The axial ratio, calculated from the x-ray cell dimensions, $a:b:c=0.8989:1:0.7016$. The unit cell contents are $Cu_{10}(PO_4)_4(OH)_8=2Cu_5(PO_4)_2(OH)_4$.

Crystal forms present are $b\{010\}$, $m\{110\}$, $n\{011\}$. The polar ratio, $p_0:q_0:r_0=0.7858:0.7046:1$. The axial ratio, calculated from the goniometric measurements, $a:b:c=0.8967:1:0.7046$.

Cleavage is {110}, perfect; fracture, irregular; hardness, 5–6; luster, adamantine to vitreous; color, dark emerald-green; diaphaneity, transparent to translucent; tenacity, brittle. The calculated specific gravity is 5.24; the measured specific gravity is 5.22.

The optical data are: $n_X=1.698$, $n_Y=1.745$, $n_Z=1.793$; 2V is close to 90°; optically (+); $X=c$.

INTRODUCTION

In the Spring of 1950, through the courtesy of Mr. G. J. Ballmer, Superintendent of Mines, a dark emerald-green, encrusting copper mineral (Fig. 1) from the Chino pit copper mine, Chino Division, Kennecott Copper Corporation, Santa Rita, New Mexico, was given to the University of New Mexico for identification. The mineral was found by Mr. William Baltosser, Mine Engineer of the Chino pit, in some fissures between two quartz monzonite dikes, 100 feet below the 6165 bench. This particular area of the pit is a highly altered section where the disseminated copper ore runs 0.5–1.0 percent. Mr. Ballmer and Mr. Baltosser believed the mineral probably was one of the more common supergene green copper minerals, and requested the mineralogy laboratory of the University of New Mexico to make the distinction. Supergene minerals are not abundant at Santa Rita. Kerr et al. (1950) report the presence of cuprite, melaconite, native copper, chrysocolla, malachite, and azurite, as well as smithsonite, in the supergene zone. No other unquestioned secondary phosphates have been reported from this locality. The occasional report of turquoise in this area has never been verified.* Apatite occurs as a primary accessory mineral in both the unaltered and altered granodiorite country rock and is the probable source of the phosphate radical in the new mineral.

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* Personal communication from Mr. Ballmer.

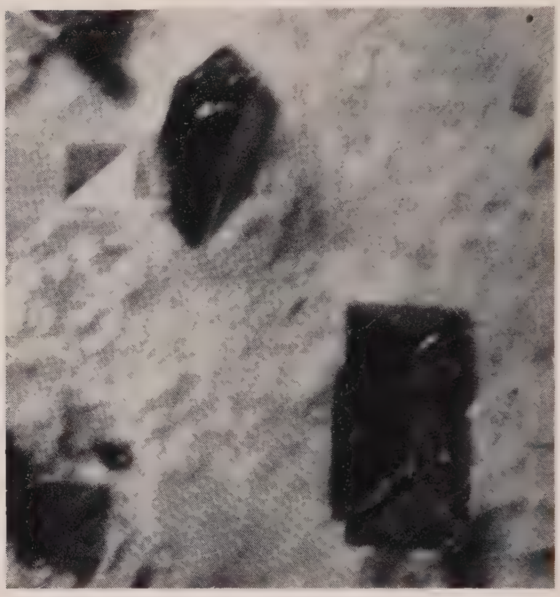

FIG. 1. Chinoite crystals on altered quartz monzonite rock. ×50.

When it became apparent, from preliminary studies, that we were dealing with a new mineral, Mr. Baltosser generously donated specimens from his private collection in order that a complete study could be made.

The writers are indebted to the late Samuel G. Gordon and Dexter H. Reynolds for scientific aid and advice given during this study.

The new mineral has been named *Chinoite* (pronounced Chee′-no-ite) after the location in which it was found.

PHYSICAL PROPERTIES

A specific gravity determination was made by means of a microchemical analytical balance, first weighing the crystal in air and then in carbon tetrachloride. This procedure was necessitated by the small size of the crystals, which range up to 1 mm. in length, and by the fact that not enough material could be sacrificed for a pycnometer determination. The crystal used in the specific gravity determinations weighed 6.26 mg. in air, 3.17 mg. in CCl_4, from which a specific gravity of 5.22 was indicated, a close check with the theoretical value, 5.24.

The cleavage of chinoite is perfect parallel to the prism {110}; fracture, irregular; hardness, 5–6; luster, adamantine to vitreous; color, dark emerald-green; diaphaneity, transparent to translucent; tenacity, brittle.

Structural Crystallography

The good crystals of chinoite were highly suitable for single crystal study. Rotation and Weissenberg x-ray pictures were taken around all three crystallographic axes using CuK_α radiation. The Weissenberg films indicated rigorously orthorhombic symmetry, and the extinctions were characteristic for the space group $Pnn—C_{2v}^{10}$. The cell dimensions, checked by the powder x-ray method, are: $a_0 = 7.47$ Å, $b_0 = 8.31$ Å, $c_0 = 5.83$ Å. The axial ratio, calculated from the x-ray cell dimensions, $a:b:c = 0.8989:1:0.7016$.

X-ray Powder Pattern

The x-ray powder pattern of chinoite is shown in Fig. 2, along with a powder pattern of pseudomalachite. The pseudomalachite powder picture was made from a sample loaned by Dr. Clifford Frondel, Harvard University (Holden Collection, #537, Rheinbreitbach, Rhine Province, Germany). The estimated intensities and the measured θ and d values for chinoite are given in Table 1. The pattern has been indexed as far as $\theta = 29.12°$, and the measured values agree with one or more theoretical values. The powder pattern of pseudomalachite was measured and calculated by Berry (1950). A comparison of the strongest lines of chinoite and pesudomalachite are given in Table 2.

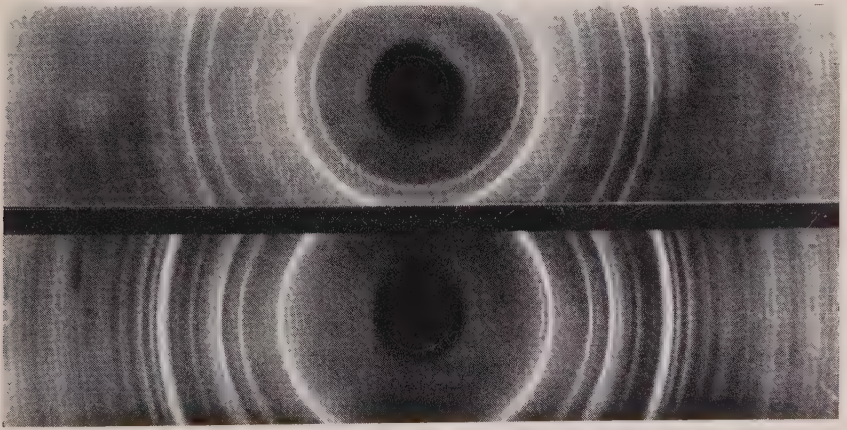

Fig. 2. X-ray powder patterns of chinoite (top) and pseudomalachite (bottom).

Morphological Crystallography

Five separate crystals of chinoite were used for the measurement of interfacial angles. Four of these crystals were mounted with the c-axis parallel to the axis of the goniometer head, and the fifth was successively

TABLE 1. CHINOITE *X*-RAY POWDER PATTERN

Orthorhombic, *Pnn*; $a=7.47$ Å, $b=8.31$ Å, $c=5.83$ Å

I	θ(Cu)	*d* (meas.)	(*hkl*)	*d* (calc.)	*I*	θ (Cu)	*d* (meas.)	(*hkl*)	*d* (calc.)
8	7.59°	5.83 Å	(001)	5.83 Å	1	24.32°	1.87 Å	(113)	1.86 Å
10	9.22	4.80	(011)	4.77				(400)	1.87
1	10.66	4.16	(020)	4.16				(141)	1.89
7	11.89	3.74	(200)	3.74	1	25.25	1.80	(210)	1.82
8	15.32	2.91	(002)	2.92				(312)	1.81
			(211)	2.88				(240)	1.82
9	16.98	2.63	(130)	2.60	2	26.68	1.71	(123)	1.71
3	17.59	2.55	(112)	2.57				(411)	1.70
5	18.49	2.43	(022)	2.39				(331)	1.73
4	19.42	2.31	(202)	2.30				(241)	1.72
			(131)	2.36				(042)	1.69
			(301)	2.29				(420)	1.70
			(330)	2.33	2	27.46	1.67	(213)	1.68
1	21.74	2.08	(231)	2.06	2	28.30	1.62	(051)	1.60
			(040)	2.08				(150)	1.62
2	23.60	1.92	(013)	1.89	2	29.12	1.58	(223)	1.58
			(132)	1.94				(402)	1.57

I	θ (Cu)	*d* (meas.)	*I*	θ (Cu)	*d* (meas.)
1	29.72	1.55	1	35.85	1.35
1	31.35	1.51	1	36.80	1.31
1	31.90	1.48	1	37.50	1.28
1	32.45	1.46	1	38.45	1.24
1	33.92	1.43	1	44.40	1.10
1	34.85	1.38	1	45.50	1.08

TABLE 2. COMPARISON OF THE STRONGEST POWDER *X*-RAY LINES OF CHINOITE AND PSEUDOMALACHITE

Chinoite			Pseudomalachite		
I	θ(Cu)	*d*(meas.)	*I*	θ(Cu)	*d*(meas.)
8	7.59°	5.83 Å	10	9.92°	4.48 Å
10	9.22	4.80	5	12.87	3.46
7	11.89	3.74	6	18.55	2.42
8	15.32	2.91	8	18.85	2.39
9	16.98	2.63	5	19.38	2.32
5	18.49	2.43	5	20.19	2.23
4	19.42	2.31	5	26.5	1.73

mounted with each of the three crystallographic axes parallel to the axis of the head. The faces were macroscopically excellent and gave good to poor optical reflections. Figure 3 illustrates a typical crystal of chinoite. Table 3 summarizes the goniometer data. The average of the measured angles agrees closely with the calculated angles. The axial ratio calculated from the goniometric measurements is in agreement with the axial ratio calculated from the x-ray cell dimensions.

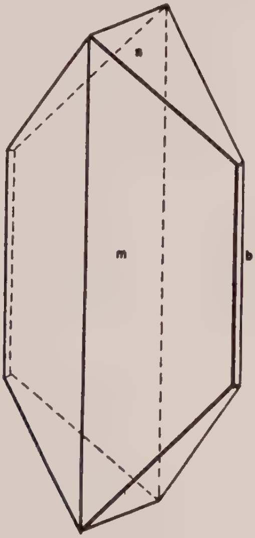

FIG. 3. Chinoite, clinographic projection showing the forms $b\{010\}$, $m\{110\}$, and $n\{011\}$.

TABLE 3. CHINOITE: CRYSTAL MEASUREMENTS, FIVE CRYSTALS

$a{:}b{:}c = 0.8967{:}1{:}0.7046$ $p_0{:}q_0{:}r_0 = 0.7858{:}0.7046{:}1$

Form	Number of Faces	Quality	Average Measured		Range		Calculated	
			Phi	Rho	Phi	Rho	Phi	Rho
$b\{010\}$	6	Poor to Fair	0°00′	90°00′	—	—	0°00′	90°00′
$m\{110\}$	18	Fair to Good	48 07	90 00	47°50′–48°16′	—	48 03	90 00
$n\{011\}$	10	Fair to Good	0 00	35 10	—	34°49′–35°25′	0 00	35 03

COMPOSITION AND CELL CONTENT

A chemical analysis of the mineral chinoite is given in Table 4. The chemical analysis gives the empirical formula, $Cu_5(PO_4)_2(OH)_4$ or $2Cu(OH)_2 . Cu_3(PO_4)_2$. Thus chinoite is dimorphous with pseudomalachite (Berry, 1950). The chemical analysis combined with the measured specific gravity (5.22) and the cell dimensions (Table 1) gives the number of molecules in the unit cell, namely, two, from which the following structural formula for chinoite is obtained:

$$Cu_{10}(PO_4)_4(OH)_8 = 2Cu_5(PO_4)_2(OH)_4.$$

The specific gravity calculated for this structural formula is 5.24.

TABLE 4. CHINOITE: COMPOSITION AND CELL CONTENT

Constituent	Percent	Molecular Ratio	Empirical Formula	Ideal Cell Content
CuO	69.09	5.025	$Cu_5(PO_4)_2(OH)_4$	
P_2O_5	24.33	1.000	or	2
H_2O	6.57	2.068	$2Cu(OH)_2 . Cu_3(PO_4)_2$	
	99.99			

0.2399 gm. analyzed. Matthew E. Carlisle, analyst.

OPTICAL PROPERTIES

The optical properties of chinoite, as determined by the immersion method, are: $n_X = 1.698$, $n_Y = 1.745$, $n_Z = 1.793$; 2V is close to 90°; optically (+); $X = c$. The optical properties of verified pseudomalachite are cited by Berry (1950) as: $n_X = 1.785 - 1.80$, $n_Y = 1.835 - 1.86$, $n_Z = 1.845 - 1.88$; $2V = 46° - 50°$; $c \wedge X = 21° - 23°$.

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KERR, P. F., KULP, J. L., PATTERSON, C. MEADE, AND WRIGHT, ROBERT J. (1950), Hydrothermal alteration at Santa Rita, New Mexico: *Bull. Geol. Soc. Am.*, **61,** 275–347.

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PREHNITIZATION OF ALBITITE*

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ABSTRACT

A steeply dipping albitite dike, 25 feet wide, which cuts serpentine is exposed in a quarry about one-half mile west of Gold Bridge in southwestern British Columbia. Prehnite has replaced part of the dike, especially its footwall where a conspicuous white irregular layer averaging 5 feet in width consisting mainly of prehnite has been formed. A comparison of the chemical analysis of this highly altered rock from the footwall (68% prehnite, 31% diopside, 1% leucoxene, epidote, and quartz) with that of less altered rock from the center of the dike (77% albite, 15% prehnite, 4% chlorite, 3% quartz, 1% leuxocene, epidote, etc.) shows principally an increase from 6.32 to 26.34% CaO and decreases from 4.93 to 0.09% Na_2O and from 57.38 to 44.84% SiO_2. It is possible that this metasomatism has been caused by fluids containing lime yielded by the serpentinization of clinopyroxene-bearing peridotite.

INTRODUCTION

Gold Bridge is approximately 100 miles north of Vancouver on the eastern flank of the Coast Mountains in southwestern British Columbia (Fig. 1). A large quarry near Gold Bridge, from which fill was obtained for the Lajoie Dam on the Bridge River, exposes a contact between ser-

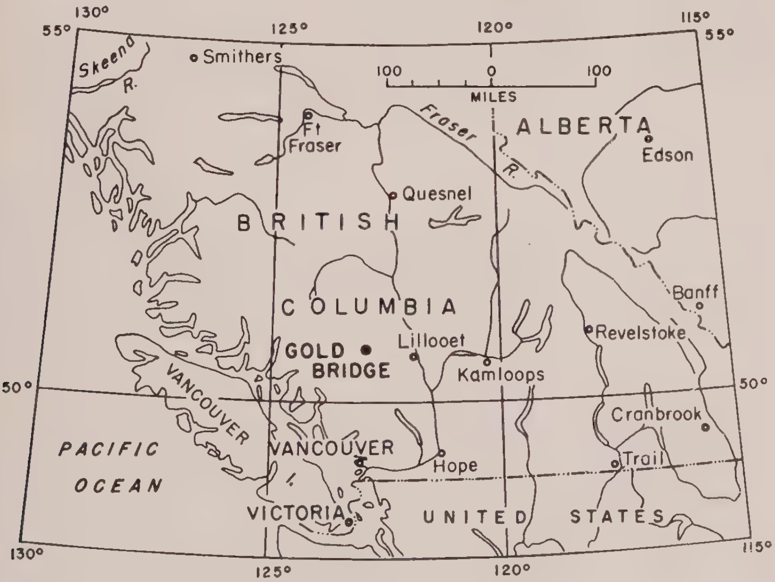

FIG. 1. Index map of southern British Columbia showing position of Gold Bridge.

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* Presented April 8, 1950, at the meeting of the Cordilleran Section, Geological Society of America held in Seattle, Washington.

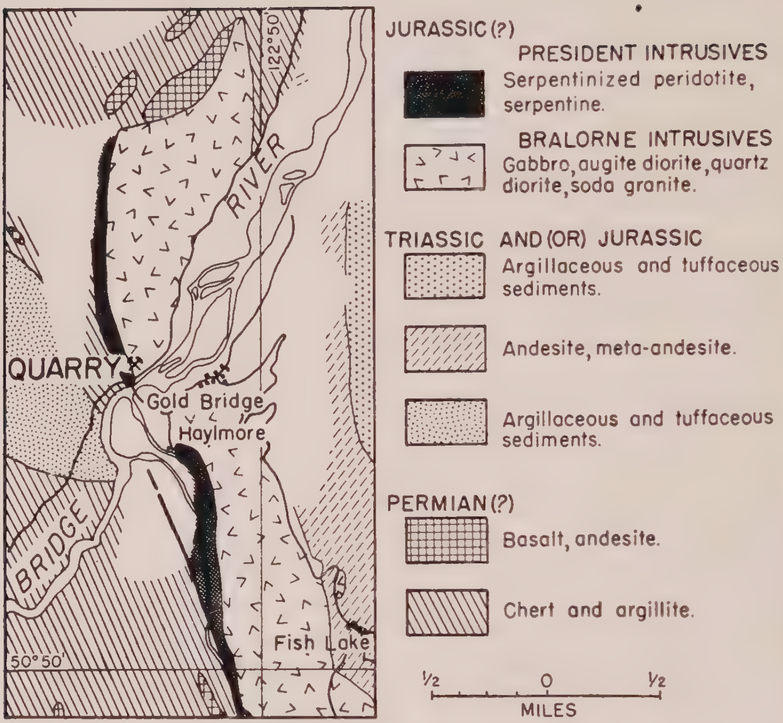
122°50'
RIVER
QUARRY
Gold Bridge
Haylmore
BRIDGE
Fish Lake
50°50'
JURASSIC (?)
PRESIDENT INTRUSIVES
Serpentinized peridotite, serpentine.
BRALORNE INTRUSIVES
Gabbro, augite diorite, quartz diorite, soda granite.
TRIASSIC AND (OR) JURASSIC
Argillaceous and tuffaceous sediments.
Andesite, meta-andesite.
Argillaceous and tuffaceous sediments.
PERMIAN (?)
Basalt, andesite.
Chert and argillite.
½ 0 ½
MILES

Fig. 2. Geological map of an area in the vicinity of Gold Bridge (after C. E. Cairnes).

pentine of the President intrusives and gabbro of the Bralorne intrusives (Fig. 2). Near this contact, which is at the western end of the quarry, an albitite dike striking about N. 10° E., dipping about 70° E., and having an average width of about 25 feet, intrudes the serpentine (Fig. 3). Much of this dike has been partly replaced by prehnite and large parts of it, especially along its footwall side, have been completely replaced by prehnite and minor diopside.

Albitite

The least altered albitite, which occurs near the center of the dike, is a light gray to light greenish gray, massive, aphanitic rock in which a very few grains of feldspar and chlorite are recognizable with a hand lens. Microscopically, it is seen that the rock is composed mainly of subhedral tabular crystals of albite, 0.1 to 0.3 mm. long, which have a tendency toward parallel orientation in places (Fig. 4). Rarely, albite crystals reach 0.5 mm. in length. The composition of the albite, determined by measurement of three grains by Universal Stage, is An_3, the range in determinations being from An_2 to An_4. Minor constituents of the albitite from the center of the dike are chlorite, quartz, leucoxene, and epidote occurring interstititally to the albite, and prehnite occurring as a replacement of some of the albite and chlorite (Fig. 4). A small amount of a mineral with radial fibrous habit, which may be anthophyllite, accompanies the prehnite locally.

A chemical analysis of the least altered albitite is given in Table 1 and the mode of the analyzed rock, determined by Rosiwal analyses of five thin sections, is included in the table.

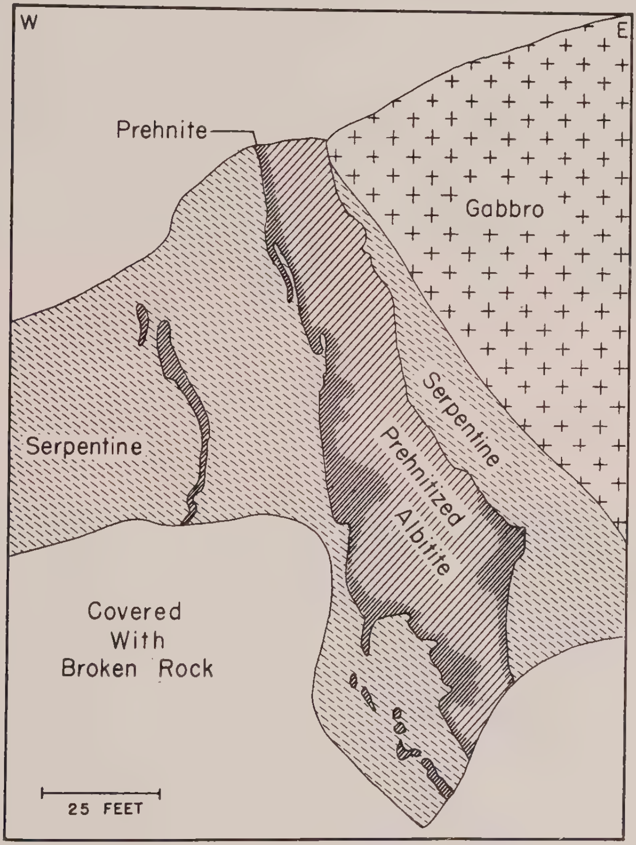

Fig. 3. Section showing geology of the face near the western end of the quarry.

Except for the alteration to prehnite and diopside, this dike is similar to many other albitite and quartz albitite dikes in the Bridge River mining district which have been described by Cairnes (1937, pp. 34–36). In

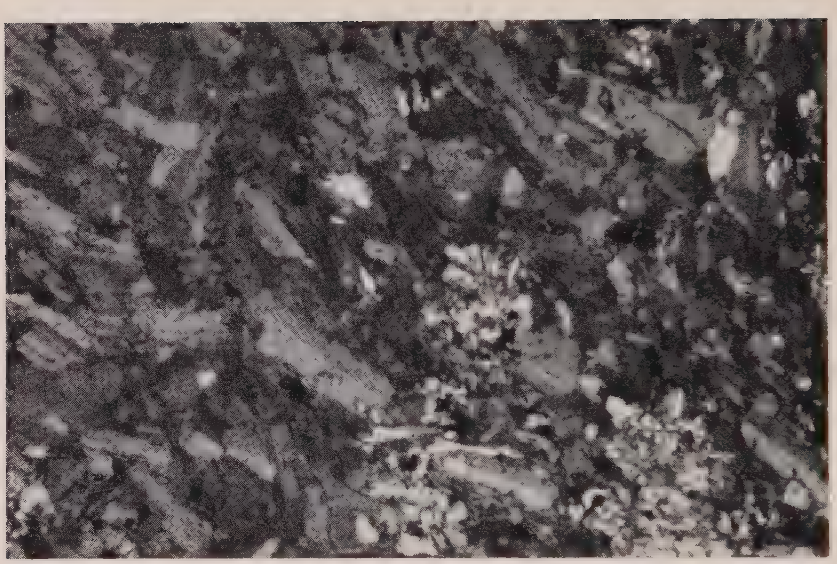

Fig. 4. Albitite showing tendency toward parallel orientation of subhedral tabular crystals of albite. Prehnite (white) has partly replaced albite. Crossed nicols. ×100.

several other places in the district, the President intrusives are intersected by these dikes (Cairnes, 1937, p. 31). A full discussion of the origin of the albitite is beyond the scope of this paper. There is little doubt, however, that because of its abrupt contacts, this dike is intrusive into the serpentine and there is no evidence that it formed by the replacement *in situ* of some pre-existing rock.

Prehnitized Albitite and Prehnite-Diopside Rock

The highly altered part of the albitite dike, which is mainly along its footwall contact (Fig. 3), is a conspicuous white rock composed of fine-grained prehnite or prehnite and diopside. The contact between the prehnite rock and the albitite is gradational; indeed, prehnite occurs in appreciable amounts even in the center of the dike. In contrast, the contacts between the massive white prehnite and serpentine are very sharp.

Two irregular and broken, narrow zones of prehnite rock lie several feet west of the main dike (Fig. 3). These are believed to be an offshoot of the main dike and a small parallel dike that have been completely replaced and have been fragmented by shearing movement in the serpentine.

The highly altered albitite is a granoblastic rock of variable mineral composition and texture. Much of it consists of irregular or slightly radiating aggregates of elongate prehnite grains 0.1 to 1.0 mm. in length and of smaller xenoblastic to short prismatic crystals of diopside in lesser amounts. The higher relief and inclined extinction of the diopside make it readily distinguishable from the prehnite in thin section, but it is difficult to see in hand specimen. Commonly, prehnite occurs alone as radiating sheaves 1 to 2 mm. long (Fig. 5). In places, the diopside is idioblastic with long slender prismatic habit. Small quantities of leucoxene, epidote, and quartz are generally present and anthophyllite (?) with radial fibrous habit occurs sporadically. Commonly, the prehnite for one or two inches near the inner edges of the completely replaced albitite contains red dust, probably hematite, that gives the rock a pink color.

A chemical analysis of typical prehnite-diopside rock from the footwall of the dike and a mode based on Rosiwal analyses of five thin sections is given in Table 1.

The properties of the prehnite are:

$G = 2.88 \pm .02$
$\alpha = 1.609$, $\beta = 1.617$, $\gamma = 1.635$; all $\pm .002$
$\gamma - \alpha = 0.026$
(+) $2V = 68°$ (calculated)
(+) $2V = 69° \pm 1°$ (U. S. measurement of 6 grains).

Measurements of a grain having wavy extinction—a fairly common characteristic of the prehnite—which gave optic angles of 64°, 61°, and 59½° in different parts, were not included in the average above. The specific

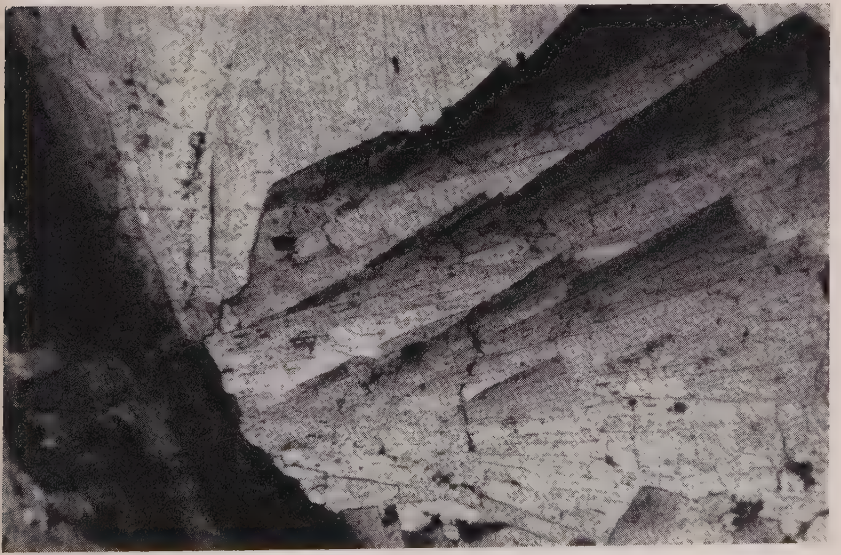

FIG. 5. Radiating sheaves of prehnite. Crossed nicols. ×90.

Table 1

Chemical analyses			Modes			
	Albitite	Prehnite-diopside Rock	Albitite		Prehnite-diopside Rock	
SiO_2	57.38	44.84	Albite	77	Prehnite	68
TiO_2	.44	.29	Prehnite	15	Diopside	31
Al_2O_3	19.96	16.32	Chlorite	4	Leucoxene, Epidote and Quartz	1
Fe_2O_3	.18	.03	Quartz	3		
FeO	5.49	5.40	Leucoxene and Epidote	1		
MnO	.22	.23				
MgO	2.14	3.32				
CaO	6.32	26.34				
Na_2O	4.93	.09	Specific Gravity	2.70 ±.01	Specific Gravity	2.96 ±.01
K_2O	tr	.07				
H_2O+	2.28	3.10				
H_2O-	.26	nil				
P_2O_5	.20	.15	Porosity	0.53% ±.01	Porosity	0.81% ±.01
CO_2	nil	nil				
Total	99.80	100.18				

W. H. Herdsman, *Analyst.*

gravity and refractive indices are slightly lower than those recorded for prehnite by Nuffield (1943, p. 62) and by Winchell (1951, p. 360).

The composition of the diopside was difficult to determine exactly because of its small grain size. The refractive indices ($\alpha = 1.690$, $\gamma = 1.715$, both $\pm .002$) and U. S. measurements ($Z \wedge c = 43°$, $(+)\ 2V = 60°$) indicate that the pyroxene is about $\frac{2}{3}$ diopside and $\frac{1}{3}$ hedenbergite (Winchell, 1951, p. 413). This composition corresponds approximately with that calculated from the chemical analysis and mode of the prehnite-diopside rock given in Table 1 and from an appropriate analysis of prehnite selected from Nuffield's data (1943, p. 62).

SERPENTINE AND GABBRO

The serpentine is a greenish black, aphanitic rock cut by closely spaced, curved, slickensided fractures coated with dark green, waxy, serpentine. Most of the rock consists of pseudomorphs after olivine which are composed of magnetite and serpentine with typical mesh structure. Each polygonal area has a rim of radially-arranged, birefringent, serpentine fibers grading inward to a core of weakly birefringent or isotropic serpentine. The rock also contains a small proportion of pseudomorphs after pyroxene which formed irregular grains interstitial to the olivine. Although a few of these pseudomorphs are typical bastite, most of them consist of serpentine flakes and fibers in various orientations. The presence of these two kinds of pseudomorphs suggests that both orthopyroxene and clinopyroxene originally occurred in the rock. Chromite is a minor accessory.

The gabbro is a somewhat brecciated, medium gray, xenomorphic granular rock consisting of about equal amounts of plagioclase and uralitized augite. Its grain size is distinctly variable though it is mostly medium. About one-fourth of the plagioclase (An_{80}) is replaced by sericite and clinozoisite and about one-third of the augite is replaced by actinolite. The gabbro contains a few patches of serpentine which may have formed from olivine.

Chemical Nature of the Metasomatism

The main reasons for concluding that the prehnite-diopside rock has formed by replacement of albitite are:

1. In the partly altered albitite many of the subhedral tabular crystals of albite are irregularly embayed or completely transected by prehnite.
2. Except for the presence of the prehnite and minor diopside the dike is similar to the numerous unaltered albitite and quartz albitite dikes in the district.
3. The prehnite-diopside rock occurs at the margins of the dike and grades into almost normal albitite at the middle.

A comparison of the chemical analyses of slightly altered albitite from the middle of the dike and prehnite-diopside rock from the footwall given in Table 1 shows that the alteration consists principally of great increase of lime from 6.32 to 26.34% and loss of soda. Decrease in silica and alumina and increase in magnesia are also evident.

It is thought that the metasomatism was essentially isovolumetric because of the following observations:

1. The porosities of the slightly altered albitite and of the completely altered rock are not significantly different (Table 1).
2. The albitite adjacent to the highly altered parts of the dike is not closely fractured as it might be if the volume had increased significantly.
3. The highly altered parts of the dike are not closely fractured nor are the parts adjacent to them mashed as they might be if the volume had decreased significantly.

The nature of the change is shown by the following comparison of the standard cells of the rocks.[1]

ALBITITE

$Na_{8.6}\ Ca_{6.1}\ Mg_{2.9}\ Fe_{4.2}\ Mn_{0.2}\ Al_{21.1}\ Ti_{0.3}\ Si_{51.5}\ P_{0.2}[O_{146.4}(OH)_{13.6}]_{60}$

PREHNITE—DIOPSIDE ROCK

$K_{0.1}\ Na_{0.2}\ Ca_{26.9}\ Mg_{4.8}\ Fe_{4.3}\ Mn_{0.2}\ Al_{18.3}\ Ti_{0.2}\ Si_{42.8}\ P_{0.1}\ [O_{140.4}(OH)_{19.6}]_{160}$

TABLE 2. ALBITITE BECOMES PREHNITE-DIOPSIDE ROCK

	By adding	By subtracting
	0.1 ion of K	8.4 ions of Na
	20.8 ions of Ca	2.8 ions of Al
	1.9 ions of Mg	0.1 ion of Ti
	0.1 ion of Fe	8.7 ions of Si
	6.0 ions of H	0.1 ion of P
Total	28.9 cations	20.1 cations
	52 valences	52 valences

POSSIBLE SOURCES OF METASOMATIZING FLUID

Several possible sources of the fluid that brought lime into the albitite may be suggested. The fluid may have been formed, for example, during the uralitization of augite in the nearby basic rocks of the Bralorne intrusives or it may have been derived from the younger bodies of granitic rocks in the area which are described by Cairnes (1937, pp. 37-39). However, it may be significant that bodies of lime-bearing silicate minerals such as grossularite, idocrase, prehnite, diopside, zoisite, and epidote have been formed as veins and as replacements of other rocks *within serpentinized ultramafics* in many parts of the world (see Benson, 1914, pp. 682-687 and 1918, pp. 722-723; Cooper, 1936, pp. 31-34; Flett and Hill, 1912, pp. 115, 143; Graham, 1917, pp. 174-176; Grange, 1927; and Turner, 1933, pp. 269-271. Benson (1918) and Turner refer to many additional occurrences). Some writers have postulated that lime becomes concentrated in a residual fluid during crystallization of an ultramafic magma. Others have postulated that conversion of clinopyroxene to serpentine in some peridotites would yield lime that may subsequently enter the new lime-bearing silicates.

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[1] Barth (1952, p. 82) has found that "In most rocks oxygen makes up about 94 per cent by volume; all cations taken together (silicon and metals) make up less than 6 per cent by volume." He concludes that " . . . it is reasonable to suppose that the mechanism of . . . isovolumetric alterations is that of a migration and exchange of cations in a medium composed of relatively stationary oxygen ions" (1948, p. 50). He has proposed using as a standard of comparison in isovolumetric changes the standard cell which is defined as "a rock volume containing exactly 160 oxygen ions" (1948, p. 51). "The sum of the cations (silicon and metals) associated with this unit is very nearly 100 in all ordinary rock types" (1948, p. 50).

The latter hypothesis is considered satisfactory to explain the prehnitization of the albitite near Gold Bridge for the following reasons:

1. The rock in which the dike occurs is now completely serpentinized, but it contains possible pseudomorphs of clinopyroxene.
2. A relatively unaltered part of the same ultramafic body contains diopside.[2]
3. The most common varieties of relatively fresh ultramafics in the district are peridotites containing diallage or enstatite or both (Cairnes, 1937, p. 28).
4. The lime content of a sample of serpentinized peridotite from the area is 3.85% but the contents of two samples of serpentine are only 0.22% and 0.30% (Cairnes, 1937, p. 29).[3]

CONCLUSION

Field and microscopic observations show that the prehnite and diopside have formed by replacement of albitite. It is possible that the lime necessary for this change has been yielded by the serpentinization of clinopyroxene-bearing peridotite.

ACKNOWLEDGMENTS

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CRYSTAL CHEMISTRY AND DIFFERENTIAL THERMAL EFFECTS OF DOLOMITE*†

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Abstract

It is observed that the lower temperature endothermal effect of dolomite takes place at a considerably lowered temperature after the dolomite has been subjected to prolonged vigorous grinding.

The mechanisms and consequences of plastic deformation in the dolomite crystallization are discussed, and are contrasted with the analogous deformations in the simple carbonates. Multiply deformed dolomite crystals actually contain island nuclei locally enriched in one or the other of the positive ions. Powder diffraction diagrams of such deformed crystallites exhibit greatly attenuated intensities for those reflections for which the form factor includes the term Ca-Mg.

An approximate evaluation of the four variable parameters of dolomite (in $R\,\bar{3}$) has been made, based on the powder diffraction. 2 C are in 2(c) with $x_c=0.243$ and 6 O are in 6(f) with $x_0=0.500$, $y_0=-0.042$ and $z_0=0.271$.

Among the abnormal differential thermal features exhibited by many dolomites, it has become apparent that some, at least, may be consequences of structural factors. An attempt has been made through extended grinding to alter mechanically the structural arrangement of a dolomite lattice and to record this alteration through variations of the differential thermal pattern and the *x*-ray diffraction effects.

The nature of the material resulting from this type of mechanical stressing precludes single crystal work, making the use of powders necessary for both *x*-ray diffraction and thermal analysis. It is thus desirable to establish with moderate accuracy the variable carbon and oxygen parameters. By drawing heavily on the crystal chemistry data available for the various simple rhombohedral carbonates, a suitable model is readily established from which refinements can be made.

The dolomite structure consists of hexagonally packed layers of carbonate groups between which alternate layers of octahedral interstices are occupied respectively by calcium and magnesium. A trial model is obtained by combining one layer of $CaCO_3$ from calcite with one layer of $MgCO_3$ from magnesite in such a manner that each layer assumes the dolomite translation, and the carbonate groups of each become coincident. This operation moves the carbonate groups toward the magnesium

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† This contribution is condensed from material originally planned as separate publications by the respective laboratories.

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layers and maintains one of the carbonate oxygens in each of the three cleavage planes.

The simple rhombohedral carbonates crystallize in $R\bar{3}c$ and dolomite in $R\bar{3}$; the lattice translations are all accurately known (Bragg) (1). The best established C—O bond length is the 1.31 Å value deduced by Elliott (3). Using this value, the Ca—O bond in calcite is 2.34 Å, and the Mg—O bond in magnesite is 2.06 Å.

When set up so that planar CO_3 groups of C—O bond 1.31 Å are oriented with one oxygen in each cleavage plane, the structure is effectively reduced to the one variable parameter x_c. Best agreement between observed and calculated powder diffraction intensities is near $x_c = 0.243$. Agreement deteriorates noticeably at x_c deviations of $\pm .003$. At $x_c = 0.243$ the Ca—O and Mg—O bonds are respectively 2.34 and 2.08 Å. In Table 1, spectrometer counts from a G.E. recording spectrometer reduced according to the formula:

$$q\alpha \frac{1 + \cos^2 2\theta}{\sin^3 \theta \cos \theta} pF^2$$

and, approximately scaled to absolute F's, are compared with calculated form factors for Ca at 0, 0, 0; Mg at $\frac{1}{2}, \frac{1}{2}, \frac{1}{2}$; 2C at $\pm x_c = 0.243$; and 6O in 6(f) at $x_0 = 0.500$, $y_0 = -0.042$, $z_0 = 0.271$, etc. The factor

$$\frac{1}{\sin \theta}$$

is introduced as an arbitrary correction for focusing conditions.

The atomic scattering factors employed are those of Pauling and Sherman (8). Some improvement in agreement for the first few lines would be realized by allowance for the partial ionic character of the Me—O bonds. Further refinement of parameters would require recourse to higher angle diffraction and single crystal methods, which would differentiate reflections to which the oxygen contributions are of two kinds.

The quality of agreement of the data in Table 1 establishes that the above-described model is an adequate basis on which to discuss the mechanical processes described below.

The effects of grinding field-samples of dolomite were investigated as a function of time by placing them in an electric-powered mortar and pestle from which samples of the material being ground were withdrawn at various intervals for differential thermal analysis and *x*-ray diffraction analysis. It was soon noted that prolonged grinding produced definite abnormalities in the physical characteristics of the materials. Therefore, the experiments were duplicated with U. S. B. S. Dolomite No. 88 in an attempt to assign the abnormalities directly to the dolomite structure rather than to some unknown condition within the field sample. The

Table 1. Comparison of Calculated and Observed Relative F Values for Powder Diffraction Lines up to $2\theta = 70°$

Index	sin θ	Form Factor			F Calc.		Reduced from Spectrometer Count
100	.1243	Ca−Mg	+0.09 C	−0.33 O	+ 4.3		2.8
110	.1357	Ca+Mg	−1.99 C	−1.41 O	+ 7.4		5.0
112	.1736	Ca+Mg	+1.97 C	+1.49 O	+41.7		49
222	.1878	Ca+Mg	−1.93 C	−5.79 O	−18.4		20.0
221	.1974	Ca−Mg	+0.43 C	+0.70 O	+10.7		11.5
$1\bar{1}0$	.2084	Ca+Mg	+2 C	−2.44 O	+15.3		12.5
210	.2286	Ca−Mg	−0.26 C	+4.63 O	+28.0	23	19
120				−4.00 O	−17.3		
$11\bar{1}$	.2427	Ca−Mg	+0.09 C	+1.27 O	+11.1		11.3
200	.2486	Ca+Mg	−1.99 C	+1.80 O	+23.8		24.5
322	.2499	Ca−Mg	−0.60 C	−0.06 O	+ 2.2		2
220	.2713	Ca+Mg	+1.97 C	−2.13 O	+15.5		14
332	.2778	Ca+Mg	+1.86 C	+1.23 O	+29.7		40*
123	.2806	Ca+Mg	−1.93 C	+3.48 O	+29.6	24	30*
213				+1.23 O	+19.9		
333	.2815	Ca−Mg	+0.80 C	+2.40 O	+16.4		
311	.2870	Ca−Mg	+0.43 C	−1.74 O	− 2.5		no certain count
$2\bar{1}0$	.3199	Ca−Mg	+0.09 C	+3.10 O	+15.4	13	12
$1\bar{2}0$				−3.71 O	−10.5		
$1\bar{1}2$	.3245	Ca+Mg	−1.99 C	+0.20 O	+14	19	17
$\bar{1}12$				+2.46 O	+22.6		
334	.3353	Ca+Mg	−1.80 C	−1.48 O	+ 8.3		11
130	.3420	Ca+Mg	+1.97 C	−1.48 O	+15.9	17	17
310				−0.90 O	+18.0		
224	.3472	Ca+Mg	+1.86 C	−1.35 O	+16.2		21*
234	.3504	Ca−Mg	+0.80 C	+3.08 O	+15.5	14	16*
324				−4.95 O	−12.5		
230	.3546	Ca−Mg	+0.43 C	+3.08 O	+14.7	11	12
320				−2.70 O	− 5.4		
$11\bar{2}$	.3610	Ca+Mg	+2 C	+3.21 O	+31.5		26
300	.3730	Ca−Mg	−0.26 C	+0.18 O	+ 3.1		
$22\bar{1}$	.3731	Ca−Mg	−0.26 C	−1.02 O	− .8		36
444	.3754	Ca+Mg	+1.73 C	+5.19 O	+37.8		

Entries marked (*) include some spurious count from incompletely resolved adjacent strong lines.

same striking results, obtained with the field sample, were duplicated precisely through the grinding of the standard material.

After continuous grinding for a period of eight hours, dolomite was noted by differential thermal analysis to undergo an additional, small

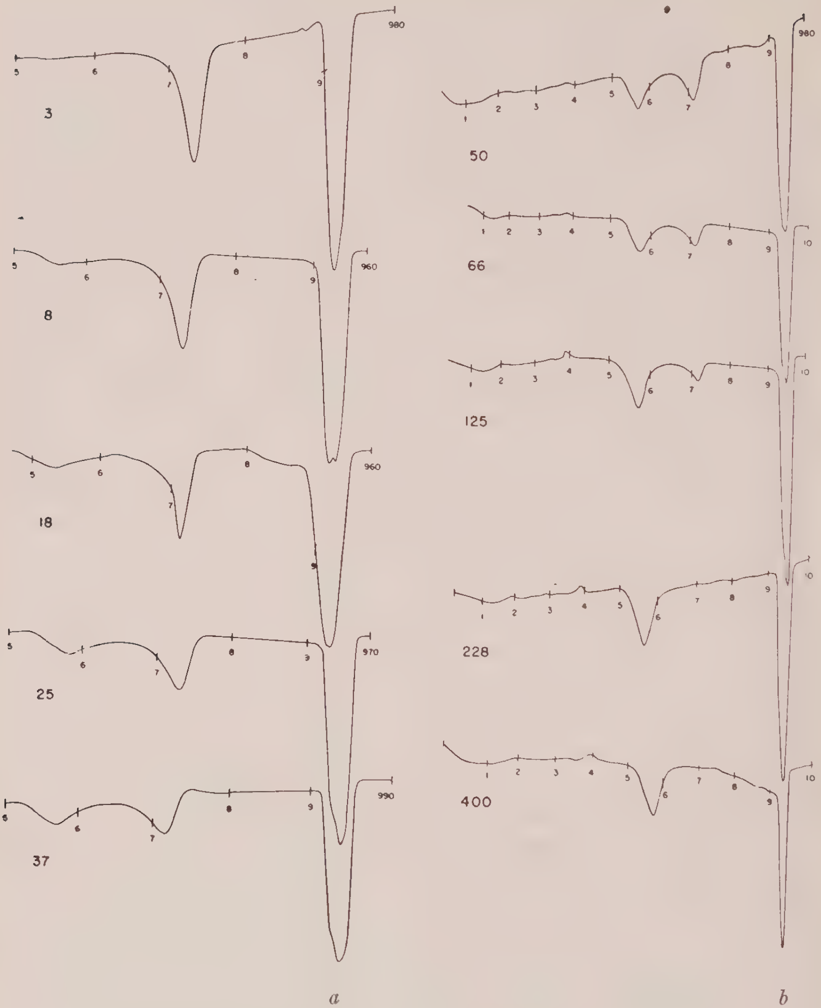

FIG. 1. Differential thermal analysis curves of U. S. B. S. Dolomite No. 88. Index numbers denote the times of grinding.

endothermic reaction between 500° and 600° C. This endothermic reaction is shown in curve 8, figure 1*a*. Curves on this plate show changes resulting from grinding, as recorded by differential thermal analysis. By employing Rowland and Lewis' (9) technique of furnace-atmosphere control, the carbonate peaks are separated and the area of each loop easily discernible. Curves in figure 1*a* show the thermal expressions from 500° to 1000° C. for material ground 3, 8, 18, 25, and 37 hours, respectively. Curves in figure 1*b* show thermal expressions for the entire range, or room temperature to 1000° C., for material ground 50, 66, 125, 228, and 400 hours, respectively. The chart speed was decreased after the sample ground for 37 hours was analyzed because it had become apparent that a new endothermic loop was being formed which would be more suitably represented in a recording made at a slower chart speed.

The continued grinding increases the magnitude of the 500°–600° C. endothermic reaction at the expense of the 700°–800° C. endothermic reaction, commonly spoken of as the evolution of CO_2 from the $MgCO_3$ portion of the dolomite composition. This reciprocal variation is illustrated by the nested curves in figure 1. Also, it was found that the ratio of the sum of the areas of the two low-temperature peaks to the area of the high-temperature peak was virtually identical in each record. Several interesting relationships between the areas of the various loops are presented in figure 2.

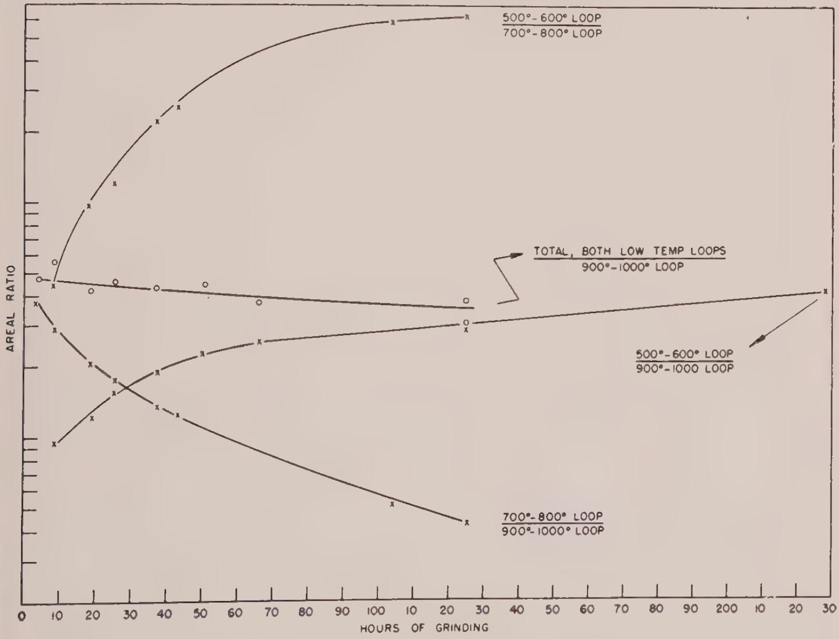

FIG. 2. Areal relationships of the three separate endothermal *DTA* features of various ground dolomite specimens.

It should be noted that the newly formed loop (figure 1) is considerably removed along the temperature axis from its progenitor and is, in this respect, unlike thermal anomalies which arise from particle-size diminution alone. When particle size is the controlling factor, a gradual lowering of the decomposition temperature usually is noted with decreasing particle size. Interest is also aroused by the fact that after the 500°–600° C. loop is fully developed and the 700° 800° C. loop completely eliminated (228-hour grind), no further significant change is recorded even after prolonged grinding (400 hours).

X-ray powder diffraction diagrams of three ground samples, irradiated and recorded simultaneously under the strictest comparable conditions, show at least three lines which are selectively reduced in intensity upon continued grinding. These lines are recorded at $d = 4.02$ Å, $d = 2.53$ Å, and $d = 2.06$ Å, and represent, respectively, the crystal planes (100), (221), and $(1\bar{1}1)$.

From Table 1 it can be noted that each of these planes ($h+k+l=$ odd) is characterized by a form factor of the type Ca minus Mg, modified by small to moderate carbon and oxygen contributions. The line-intensity diminutions are thought therefore to result from some specific structural alteration such as a glide mechanism whereby Ca^{++} and Mg^{++} ions would be migrated along certain directions within the lattice. When sufficient Ca^{++} ions have been translated to positions formerly occupied by Mg^{++} ions, so that a random distribution of these two ions is produced in the projection of the whole structure onto the poles to the planes in question, the lines representing those planes should almost disappear.

Examination of figure 3 indicates that reflections from planes (100), (221), and $(1\bar{1}1)$, which were relatively intense after three hours of grinding, became selectively less intense after 50 hours of grinding and virtually disappeared after 228 hours of grinding. It was at this point that the *DTA* curves also became stabilized (figure 1). The fact that these two circumstances prevail concurrently indicates the attainment of a new equilibrium established through the continued application of the nondirectional stress load. Crystallographic studies of plastic deformation in ionic crystals have been concerned primarily with directed loads as exemplified by Buerger's (2) analyses of the NaCl type face-centered cubes. There is no reason to assume that the translations so induced are the only ones which occur, and indeed, Buerger infers that deformations along octahedral planes are probably induced by random loading. Deformation along octahedral planes is also a frequent inference in diffraction analyses of orientation effects produced in the working of face-centered cubic metals.

The close structural analogy between the face-centered cubic alkali halides and the rhombohedral carbonates suggests that such structural consequences of deformations as could have resulted from the above grinding experiments could profitably be examined.

In order to avoid the confusion which arises from the various sets of reference axes to which rhombohedral carbonates are at times referred,

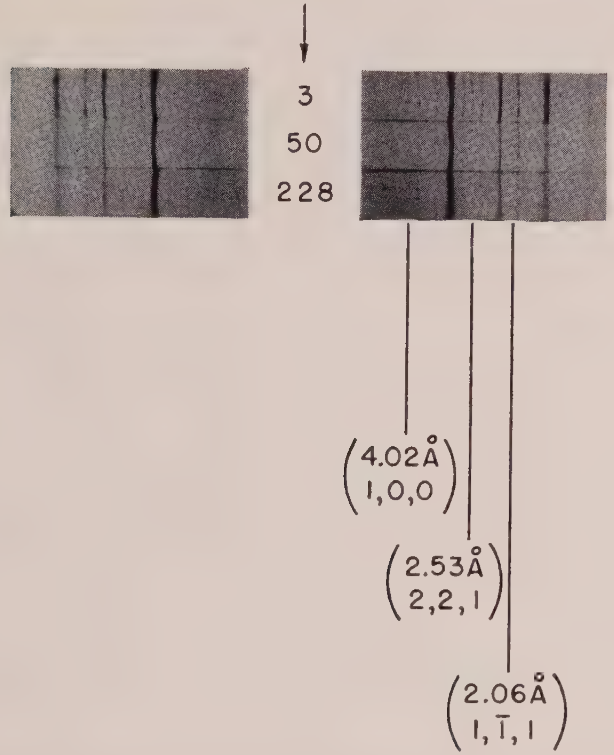

Fig. 3. Simultaneously registered *x*-ray powder-diffraction diagrams of the U. S. B. S. dolomite No. 88. Upper, after 3 hrs., middle, after 50 hrs., and lower, after 228 hrs.

Miller indices will be used in the following discussion to refer only to the true rhombohedral cell. Analogous cubic directions can then be referred to by name. Table 2 presents the various index correlations for the commoner dolomite forms.

Translation gliding of dolomite with $T\{111\}$, $t[10\bar{1}]$ was produced experimentally by Johnsen (7), and twin gliding on $\{110\}$, probably along the short diagonals of the cleavage rhomb, is discussed by Fairbairn and Hawkes (4). Both of these are octahedral glides along dodecahedral poles in the cubic system. The familiar pressure twins of calcite across (332) are not known in dolomite. This is the "forbidden" dodecahedral translation along a cube pole. It occurs in calcite only as a twin glide, not as a full translation.

Figures 4, 5, and 6 are projections of the dolomite structure onto $(1\bar{1}0)$ arranged as discussed by Fairbairn and Hawkes. Figure 4 is a projection of the aspect a calcite pressure twin would assume in dolomite if it did occur.

TABLE 2. SPECIMEN INDEX FORM CORRELATIONS FOR VARIOUS DOLOMITE REFERENCE AXES

Rhombohedral unit cell	Cleavage rhomb.*	Crystallographers hexagonal*	Hexagonal structure cell $a=4.8$ Å $c=16$ Å	Edge-centered rhomb.	Face
100	[3$\bar{1}\bar{1}$]	[40$\bar{4}$1]	10$\bar{1}$1	$\bar{1}$11	M
110	11$\bar{1}$	[02$\bar{2}$1]	01$\bar{1}$2	002	f
211	200	10$\bar{1}$1	10$\bar{1}$4	022	r
222	111	[0001]	0006	222	c
221	[33$\bar{1}$]	[04$\bar{4}$5]	01$\bar{1}$5	113	l
10$\bar{1}$	20$\bar{2}$	11$\bar{2}$0	11$\bar{2}$0	$\bar{2}$02	a
210	[51$\bar{3}$]	[44$\bar{8}$3]	11$\bar{2}$3	$\bar{1}$13	α
11$\bar{1}$	[33$\bar{5}$]	[08$\bar{8}$1]	02$\bar{2}$1	$\overline{11}$3	d
322	[511]	[40$\bar{4}$7]	10$\bar{1}$7	133	
332	220	01$\bar{1}$2	01$\bar{1}$8	224	e
321	31$\bar{1}$	[22$\bar{4}$3]	11$\bar{2}$6	024	
20$\bar{1}$	[7$\bar{1}\bar{5}$]	[8·4·$\overline{12}$·1]	21$\bar{3}$1	$\bar{3}$13	k
21$\bar{1}$	31$\bar{3}$	[24$\bar{6}$1]	12$\bar{3}$2	$\bar{2}$04	
310	40$\bar{2}$	21$\bar{3}$1	21$\bar{3}$4	$\bar{2}$24	v
2$\bar{1}\bar{1}$	4$\bar{2}\bar{2}$	30$\bar{3}$0	30$\bar{3}$0	$\bar{4}$22	m

* Brackets denote that reflections appear for fractional orders under these conventions.

When calcite is pressure-twinned the ions of each successive (332) level are displaced along a "cube" pole to the alternative equivalent position, which represents the exchange of the long for the short diagonal of the cleavage rhomb, and CO_3 ions rotate according to the steric requirements. Each successive level makes this exchange as the twin boundary

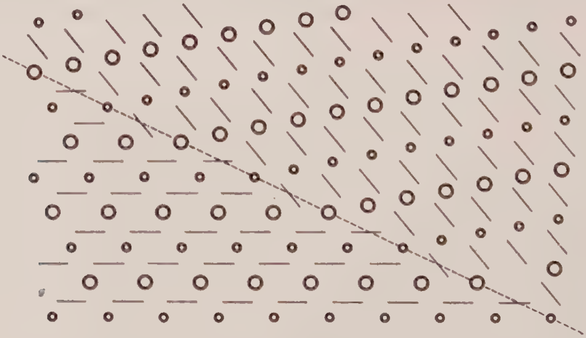

FIG. 4. A (1$\bar{1}$0) projection of the aspects which dolomite would exhibit at the twin boundary if the familiar calcite type (332) twin were produceable by pressure. Large circles represent Ca, small circles, mg, and lines, CO_3.

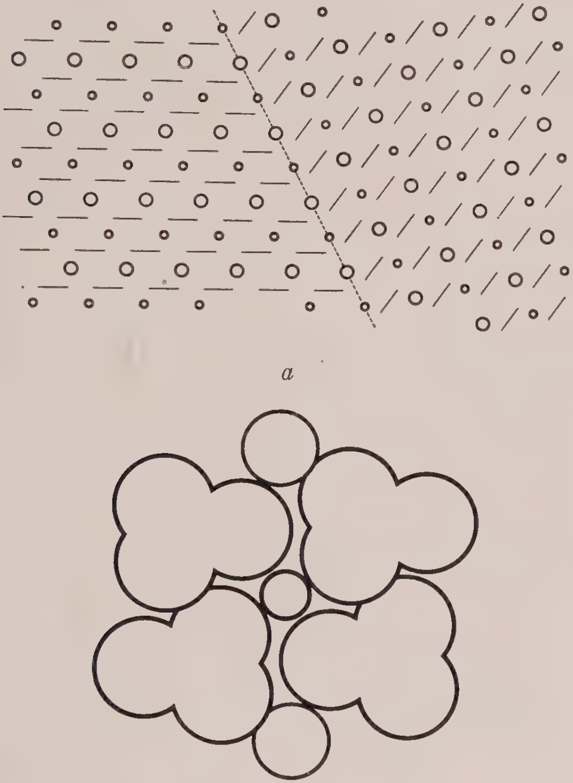

FIG. 5. A $(1\bar{1}0)$ projection (*a*) of the twin boundary of dolomite for the (110) pressure twin, illustrating the identity of the original and synthesized members. The supplemental sketch (*b*) is a view of the packing arrangement in a cleavage rhomb face.

progresses through successive 332 levels, but no level glides beyond the like-ion barrier of the "cube" pole glide. The result is that the (222) and one of the (110) directions interchange, and inasmuch as the projection of the structure upon each is a simple alternation of Ca and CO_3 ions, no drastic rearrangement has been required.

As is seen in figure 4, the application of the identical mechanism to dolomite fails as a physical possibility because of the unequal Ca and Mg ion sizes and dispositions. Projected ion sequences along [111] and [110] are, respectively, Ca, CO_3, Mg, CO_3, etc., and Ca+Mg, $2CO_3$, Ca+Mg, $2CO_3$, etc., which are not interchangeable. A pressure twin of composition face (332) would require a complete redistribution of the positive ions.

Figure 5 illustrates the apparent mechanism of the formation of the lamellar dolomite twins across (110). Not apparent in this projection, but clear in three dimensions and in the supplemental sketch, is the fact that alternating CO_3 levels as projected onto [111] have oxygen atoms pointed alternately toward and away from the possible short diagonal glide lines. Thus in gliding, a protruding oxygen atom can pass a Mg position with less friction than it can a Ca position, and a mechanism is set up to favor translations of one level only. It is seen that such unit glides (modified only by a small twin glide) along succeeding short diagonals for each successive (110) level redistribute the positive ions into dolomite successions in the synthesized twin, and simple rotation of the affected carbonate groups builds a twinned lamella. It seems reasonable that the natural twinned lamellae are so produced.

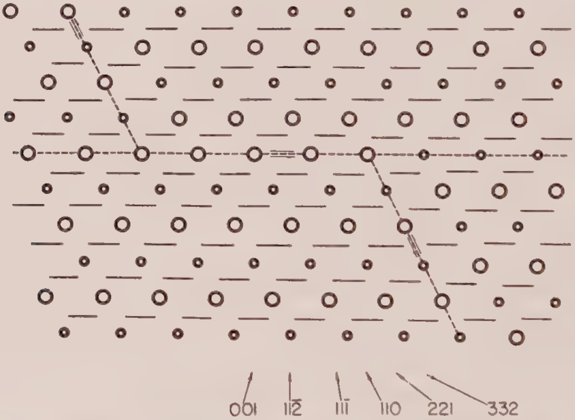

FIG. 6. A ($1\bar{1}0$) projection of a detail of a dolomite structure which has been subjected successively to a unit glide on (110) and a glide of several units on (222).

The probable mechanism of the deformation in the grinding experiments described earlier is illustrated in figure 6. The section depicts a detail of the crossing of a unit glide on (110) by a longer glide on (111). Under the varying directions and magnitudes of the loads impressed in grinding, this detail must be reproduced an enormous number of times in the reduction of natural texture dolomite rocks to impalpable powders. Each such intersection of glides produces a nucleus within the dolomite structure in which all neighbors of some carbonate groups are either all Ca or all Mg, and therefore repeated crosses can build up substantial Ca-rich and Mg-rich islands. In the same mechanism the phase relations of positive ions within coherently scattering domains are disrupted. Consequently, the diffraction lines which are peculiar to dolomite, and which derive their intensity mainly from the difference in scattering power between Ca and Mg, are attenuated. Such lines are thus distinguished from those to which the Ca and Mg make contributions of equal sign, and from those to which the major contributor is oxygen.

Haul and others (5, 6, 10) have recently discussed the thermal decomposition of dolomite and have noted the evidence for an activation energy. In consideration of the observation that calcite and periclase are produced essentially simultaneously, the calcite, at least, being synthesized at orientations controlled by the original dolomite structure, it seems reasonable to assume that the activation energy is associated with interdiffusion of positive ions to produce Ca-rich and Mg-rich localities. This kind of sorting having already been accomplished mechanically by grinding, in the case of the vigorously ground material above, thermal decomposition of this powder proceeds at a notably reduced temperature, more comparable with that at which magnesite itself decomposes.

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FLUID INCLUSIONS IN BERYL AND QUARTZ FROM PEGMATITES OF THE MIDDLETOWN DISTRICT, CONNECTICUT

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ABSTRACT

Secondary and apparently primary fluid inclusions occurring in quartz and beryl from certain pegmatites of the Middletown district, Connecticut, have been investigated by the Sorby method. In three pegmatites the inclusions found appear to be aqueous solutions of moderate concentration. In the other pegmatites most or perhaps all the inclusions contain considerable amounts of carbon dioxide. Temperature data for quartz have proved difficult to obtain, owing to fracturing and leakage of inclusions when the crystals are heated. Beryl has yielded considerable data. The data are analyzed in terms of possible physico-chemical conditions in pegmatitic liquids and in terms of the various sources of error inherent in the Sorby method. The bearing of the work on the use of liquid inclusions in determining temperatures of pegmatite mineral formation and compositions of pegmatitic liquids is discussed.

INTRODUCTION

Since 1938, field studies of pegmatites and of the occurrence of minerals in them have directed attention to a series of problems that call for laboratory studies correlated with the field investigations. The need for these studies became apparent to the first-named author during wartime investigations of pegmatite mineral deposits, and a group of beryl-bearing pegmatites in the Middletown district of Connecticut was selected for study. These pegmatites had been mapped in detail for the U. S. Geological Survey by Cameron and V. E. Shainin during the period 1942–44. In the fall of 1945, Cameron made further studies of the pegmatites in the field and collected suites of specimens. Laboratory study was undertaken at the University of Wisconsin in the fall of 1949.

The present paper is a report of the first phase of the study, during which fluid inclusions in beryl and quartz from various pegmatites have been investigated. The chief objects were to learn whether the fluid inclusion method can be used to determine temperatures of mineral formation in this group of pegmatites, and to obtain a check on temperature data of formation reported for liquid inclusions in beryl and quartz from two of the pegmatites. The investigation has also yielded information that may bear on the composition and physical characteristics of the fluids from which pegmatites crystallize.

ACKNOWLEDGMENTS

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Previous Work

Certain pegmatites of the Middletown district have been known for many years as sources of commercial feldspar and mica, and as prolific sources of mineral specimens. There are numerous articles dealing with the pegmatites; a few are especially pertinent to the present investigation.

Jenks (1935) described the results of a detailed mineralogical study of the Strickland-Cramer pegmatite. During 1942–45, many of the pegmatite bodies of the district were mapped in detail and studied by E. N. Cameron and V. E. Shainin; these studies are described in a forthcoming report. A brief description of the more important beryl-bearing pegmatites has been given by Cameron and Shainin (1947).

In 1947, Ingerson gave temperature data for liquid inclusions in beryl and quartz in specimens furnished by Cameron from the Case No. 1 and No. 2 pegmatites. Ingerson reported that bubbles in inclusions in several sections of quartz and one of beryl disappeared at temperatures ranging from 153° C. to 175 °C. Quartz from one of the specimens studied by Ingerson was subjected to decrepitation by H. S. Scott (1948, p. 650), who obtained 165° C. as the temperature of formation of the quartz (uncorrected for pressure).

Description of the Pegmatites

Location and geologic setting: The pegmatites presently being studied are mostly in a narrow strip east of the Connecticut River and parallel to it, in the vicinity of Middletown (Fig. 1). They are essentially lenticular bodies of various sizes enclosed in the Bolton schist and the Monson (Glastonbury) gneiss. One body, the Slocum pegmatite, lies east of the others, in the town of East Hampton. It occurs in interlayered schists and schistose quartzites known as the Hebron gneiss (Gregory and Robinson, 1907). The metamorphic rocks of the area are involved in a series of major folds. Age determinations from radioactive minerals obtained from several pegmatites of the district suggest that the pegmatites are of late Devonian age (Foye and Lane, 1934), and the enclosing rocks are generally considered to be of Paleozoic age.

There is no direct evidence of the thickness of rocks that has been removed since emplacement of the pegmatites. Indirect evidence suggests that the thickness may have been great. The area forms part of the Eastern Highland of Connecticut. The Triassic rocks of the Central Connecticut Lowland rest on an erosion surface that presumably extended eastward and truncated the folded, crystalline rocks of the Eastern Highland. Erosion of the Eastern Highland contributed much of the 10,000 to 15,000 feet of sediments comprising the bulk of the Triassic formations, but how much the surface developed on the crystalline rocks was lowered during that time is a matter of conjecture. An erosion surface correlated by D. W. Johnson (1931) with the Schooley peneplane truncates both the Triassic rocks and the rocks of the Eastern Highland. Furthermore, if Johnson's interpretation of the geomorphic history is correct, this surface truncates the earlier Fall Zone peneplane, on which the coastal plain sediments of Long Island rest. The history of the region since the time of folding and metamorphism therefore includes two and possibly three periods of prolonged erosion. An estimate of 15,000 feet for the minimum thickness of covering rocks removed since emplacement of the pegmatites

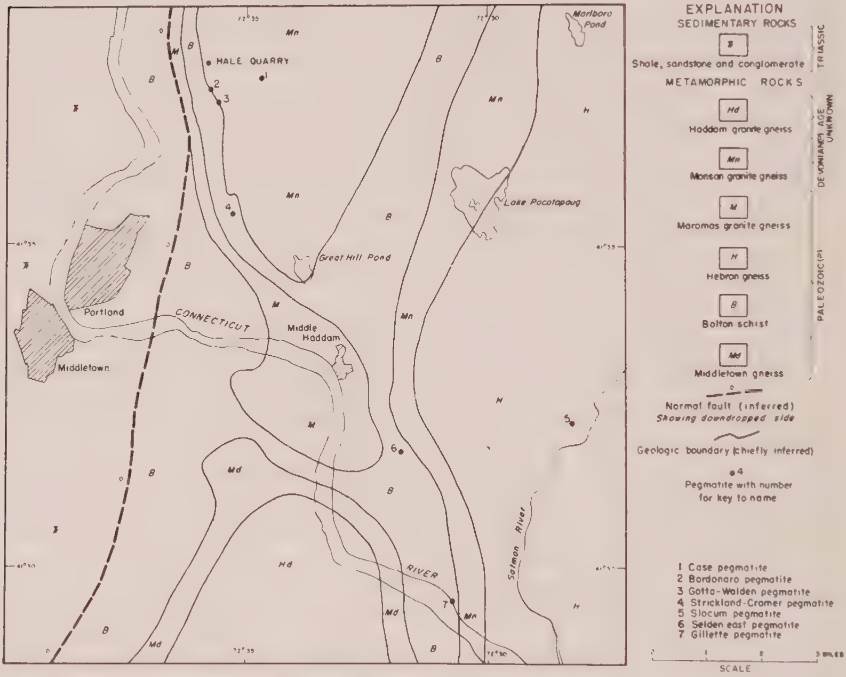

Fig. 1. Index map of the Middletown district, Connecticut, showing locations of pegmatites studied.

would therefore seem conservative. The true thickness may be considerably greater.

Character of the pegmatite bodies: The pegmatite bodies of the area range from those that are essentially uniform in mineral composition and mineral proportions to those in which zonal structures are developed in various degrees. The pegmatites that form the basis of the present investigation are a selected group, consisting largely of bodies in which zonal structures are at least moderately well developed. The sole exception is the Selden east pegmatite, in which an indistinct wall zone and pods of quartz rimmed by coarse feldspar and muscovite are the only indications of zonal structure.

The pegmatites are mostly simple in composition. The essential minerals are perthite, sodic plagioclase, quartz, and muscovite or biotite, or both. Garnet, tourmaline, and beryl are accessory minerals in all. The many other minerals found from time to time in the pegmatites are trace components.

The zones and corresponding mineral assemblages of the pegmatites are summarized in Table 1.

Table 1. Zones and Mineral Assemblages of the Pegmatites

Pegmatite	Border zone	Wall zone	Intermediate zone	Core or apparent core	Replacement bodies
Case No. 1	*P*, *Pl*, *Q*, B			*P*, *Q*, B, G, M	*Pl*, *Q*
Case No. 2	*P*, *Pl*, *Q*, M, G, B, T	*P*, *Pl*, *Q*, M, B	*P*, *Q*, B	*Q*	
Case No. 3	*P*, *Pl*, *Q*, M, G, B, T	*P*, *Pl*, *Q*, M, B, T	*P*, *Q*, Pl, B	Q (pods)	
Bordonaro south	*Pl*, *P*, *Q*, B, M	*P*, *Q*, M, Pl		Q	
Bordonaro north	*Pl*, *P*, *Q*, M, B	*Pl*, *Q*, *P*, M, B, T, G		*P*, *Q*, Pl, M, B, G	*Q*, *Pl*
Gotta-Walden	*Q*, *Pi*, M, G, T, B	*Pl*, *M*, *Q*, B, G, T		*Pl*, *Q*, *P*, *M*, B, G, T	
Selden east	*Pl*, *Q*, *M*, B	*Pl*, *Q*, *M*, P, B, G, T		*Q*, *Pl*, *P*, M, B	
Strickland-Cramer	*Q*, *M*, *Pl*, G, T	*Pl*, *Q*, *M*, T, G	(1) *Q*, *P*, *Pl*, M, G, T (2) *Q*, *Pl*, G, T	*Q*, Pl	*Pl*, *Q*, B, M, Li minerals
Slocum	*Q*, *P*, *Pl*, G, T, B	*Q*, *P*, *Pl*, B, T, M, Bi	*P*, *Q*, B	*Q*, P	

Symbols: P, perthite; Pl, plagioclase; Q, quartz; M, muscovite; Bi, biotite; G, garnet; B, beryl; T, tourmaline. Symbols for essential minerals italicized. Trace minerals omitted.

OCCURRENCE OF BERYL AND QUARTZ IN THE PEGMATITES

Beryl is unevenly distributed in most of the pegmatites. It is abundant in certain zones or replacement bodies but rare or absent in others. In the Strickland pegmatite, for example, visible beryl is largely restricted to replacement bodies, whereas in the Case No. 2 and Case No. 3 pegmatites it is concentrated in the border and wall zones. In the Gotta-Walden pegmatite, however, beryl is found in approximately equal amounts in all three zones of the body.

Beryl occurs in all pegmatites as subhedral to euhedral prismatic crystals. The crystals are 6 inches or less in length and 2 inches or less in diameter in most of the pegmatites, but larger crystals have been found in a few. In inner zones of the pegmatites, the crystals are commonly haphazardly oriented, but in outer zones beryl generally forms crystals oriented perpendicular or subperpendicular to the walls. These crystals are commonly tapered. The outer ends (nearest contacts of pegmatite with wall rock) are slender, whereas the inner ends (toward the cores of the pegmatites) are thicker. In addition, the beryl crystals in some of the pegmatites are progressively larger from zone to zone from the walls inward.

Beryl varies in color from pegmatite to pegmatite or even within the same pegmatite. Beryl from the Bordonaro, Gotta-Walden and Case pegmatites is medium green. Beryl from the Selden east pegmatite is pale green, nearly white. Beryl collected at the Slocum mine is golden, although beryl of other colors is said to have been found there. Beryl from the Strickland-Cramer mine ranges from blue to golden to pale pink. The color variations presumably correspond to variations in alkali content.

Quartz occurs in all zones and replacement bodies of all the pegmatites. It occurs mostly as anhedral grains and aggregates of anhedral grains that are interstitial to other minerals. It ranges from clear to milky to smoky in color.

APPARATUS USED FOR TEMPERATURE DETERMINATIONS

A. Apparatus for heating experiments

The apparatus used for heating experiments and the procedure for preparation of crystal plates are essentially the same as that described by Bailey and Cameron (1951, p. 634). For high-temperature work, however, a cell of smaller diameter was found more suitable. It has also been found best to polish only the upper side of the crystal plate. The lower side is ground with M-303 (American Optical Co.) emery. This ground surface provides a diffuse illumination against which inclusions can be seen more clearly through the polished upper side of the plate.

In practice, the polished plate is examined carefully under the microscope before heating runs are attempted. Liquid inclusions present are studied and classified into groups according to shape, distribution, and arrangement with respect to crystallographic features. Each group is studied separately. The approximate liquid-vapor ratios shown by the inclusions are noted. A series of inclusions is then selected for heating trials. Sketches are made showing the individual inclusions in each field of view that will be watched during heating. The inclusions are numbered, so that effects of heating can be recorded separately for each inclusion.

Heating is done at a controlled rate, in order to prevent a lag of the temperature of the plate with respect to that of the cell. It has been found that a rate of 10–15° C. per minute is satisfactory. The inclusions are watched throughout the heating process. Temperatures of vapor disappearance or of phase boundary disappearance are read by means of a potentiometer and thermocouple. For actual reading, the tip of the thermocouple is touched to the lower side of the crystal plate. The potentiometers used are calibrated against the melting points of standard alloys. Experience indicates the temperature values given in the present paper for temperatures of disappearance of a vapor phase or of a liquid phase are accurate to within $2\frac{1}{2}$° C. Values for disappearance of phase boundaries are less accurate and may be as much as 10° C. in error; the precise point at which disappearance is complete is difficult to determine.

The end result of heating most fluid inclusions is disappearance of the vapor phase. When this has been accomplished, heating is discontinued and the cell allowed to cool. The temperature of reappearance of the vapor phase is observed as a rough check on the temperature of disappearance. After cooling, each run is repeated as a check on the first determination and as a check against leakage.

Disappearance of the liquid phase or of a phase boundary is similarly checked by repeated measurements.

Many of the inclusions examined contain a second liquid phase that disappears when heated to a temperature close to 31° C. We infer this to be liquid carbon dioxide. This second liquid phase is easily overlooked, because the heat of a strong lamp may raise the temperature of inclusions above 31° C. As a check, the plate must be cooled by moistening it with acetone or alcohol and watching the results under the microcsope.

Various combinations of 10× and 20× hyperplane objectives with 10× achromatic, 20× achromatic, and Leitz UM 4 objectives have been used in the present study. The UM 4 objective combines large working distance with high magnification (30×) and is especially useful for study of minute inclusions. A water-cooled sleeve device serves to prevent heating of the objective when in use.

B. *Apparatus for freezing experiments*

Results of heating experiments raised various questions as to the composition of the liquid inclusions studied. Freezing experiments have therefore been undertaken. A simple wooden cell has been used for this purpose. Solid carbon dioxide or a current of air passed over liquid air contained in a thermos jar has been used as the coolant. With liquid air, temperatures as low as −120° C. have been reached, but condensation of moisture in the cell has caused trouble, and a dehydration device is needed.

Subzero temperatures are read by means of a thermocouple connected to a Leeds and Northrup factory-calibrated potentiometer. The temperature within the cell can be controlled within 1 or 2 degrees C. by regulating the flow of air over the liquid air.

RESULTS OF THE STUDY OF FLUID INCLUSIONS

General statement: It was intended originally to study fluid inclusions in quartz from all nine pegmatites listed in Table 1, but it soon became apparent that quartz in these pegmatites is ill suited to temperature studies. One reason is that primary and secondary inclusions in quartz are difficult to distinguish. Most of the liquid inclusions that are not clearly secondary occupy irregular cavities. The grains are anhedral and unzoned, hence the distribution of inclusions with relation to crystallographic elements is difficult or impossible to determine. The second reason is that quartz shows a marked tendency to fracture when heated, and most liquid inclusions leak before temperatures of disappearance of the vapor phase are reached. Reliable temperature determinations for inclusions in quartz are therefore very difficult to obtain. Beryl, on the other hand, occurs in well-formed crystals. The distribution of inclusions with respect to crystallographic elements is in general easily determined. Many inclusions occupy oriented negative crystal cavities or tubular cavities that show no relation to secondary structures of the crystals, and there is a strong possibility that at least some of these can ultimately be shown to be primary. Further, the shapes of many inclusions are such that during heating the changes in the volumes of the component phases can be determined accurately. Finally, the crystals show little tendency to fracture, and leakage of inclusions is not a serious problem; the majority of inclusions can be heated to temperatures 50° C. or more above the filling temperatures. Beryl is therefore very satisfactory for temperature studies and has received our principal attention.

Fluid inclusions in beryl: More than 2,000 fluid inclusions in beryl from the various pegmatites have been studied in the course of the present investigation. Temperatures of disappearance of the vapor phase, the liquid phase, or the vapor-liquid phase boundary have been determined for some hundreds of fluid inclusions. The majority of the determinations are precise measurements, inclusion by inclusion. In certain beryl crystals from the Selden, Case No. 2, and Strickland-Cramer pegmatites, however, clusters of inclusions could be observed simultaneously during heating. Precise disappearance temperatures sufficient to represent the range for each cluster have been made, and the temperatures of disappearance for the remainder have been determined approximately.

Fluid inclusions found in crystals of beryl are of 3 principal types:

(1) Disseminated negative crystals and tubular inclusions with long axes parallel to the *c*-axis of beryl (Figs. 2 and 3). These are referred to below as oriented negative crystals. In an earlier paper (Cameron, Rowe, and Weis, 1951) the tubular inclusions were regarded as a separate type, but experience has shown that a sharp line between tubular inclusions and negative crystals can-

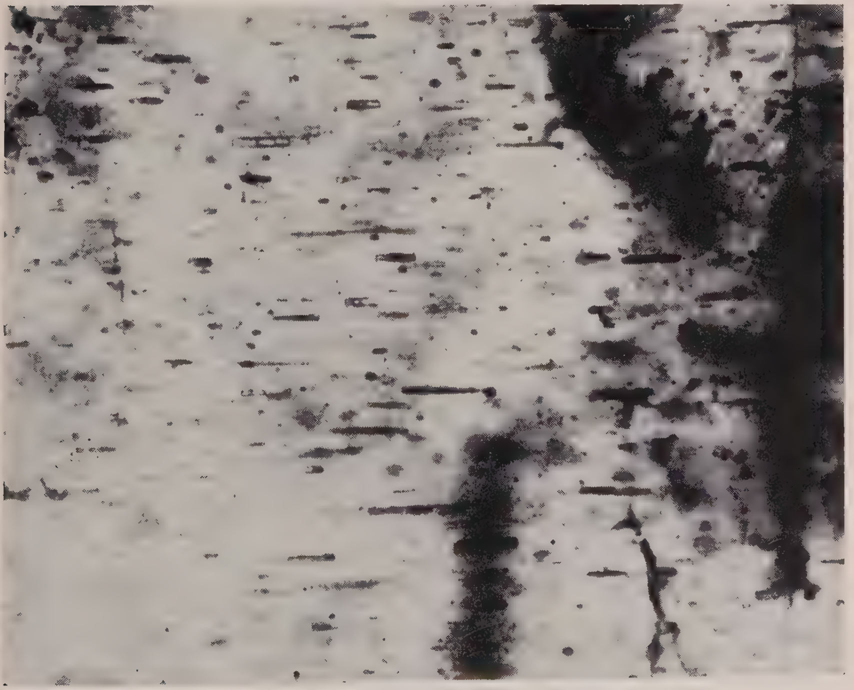

FIG. 2. Negative crystal and tubular inclusions in beryl, oriented parallel to the *c* axis. Case No. 2 pegmatite. ×200.

not always be drawn. Both may occur in the same mineral plate.

(2) Disseminated rounded, elongate, or irregular inclusions, unoriented with respect to crystallographic elements (Fig. 4).

(3) Rounded, elongate, or irregular inclusions and negative crystals arranged along planes or curving surfaces of random orientation.

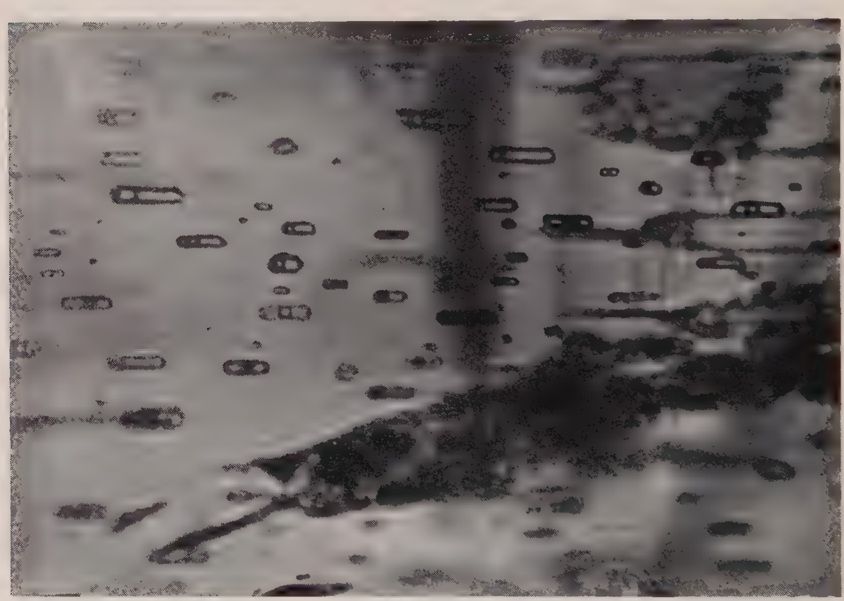

FIG. 3. Negative crystal inclusions in beryl, oriented parallel to the *c* axis. Selden east pegmatite. ×200.

Intersections of planes or surfaces are common (Fig. 5). In most of the plates cut these are the only inclusions. In most of the remaining plates, they are the most abundant type of inclusion.

Inclusions of type (3) are definitely secondary. They are distributed along healed fractures unrelated to crystal boundaries or crystallographic elements of the host beryl. The origin of the other types of inclusions is less certain. Those of type (2) show no relation either to crystallographic elements or to secondary structures. They may be either primary or secondary. Inclusions of type (1), on the other hand, are crystallographically oriented and appear unrelated to fractures or other discontinuities within the crystals. Inclusions of this type evenly disseminated in swarms in the centers of unfractured crystals are especially difficult to account for as secondary inclusions. They were clearly developed prior to the inclusions of type (3), which have a pattern that is superimposed upon the pattern of inclusions of type (1). In this connection, a large crystal of golden beryl from a replacement body in the

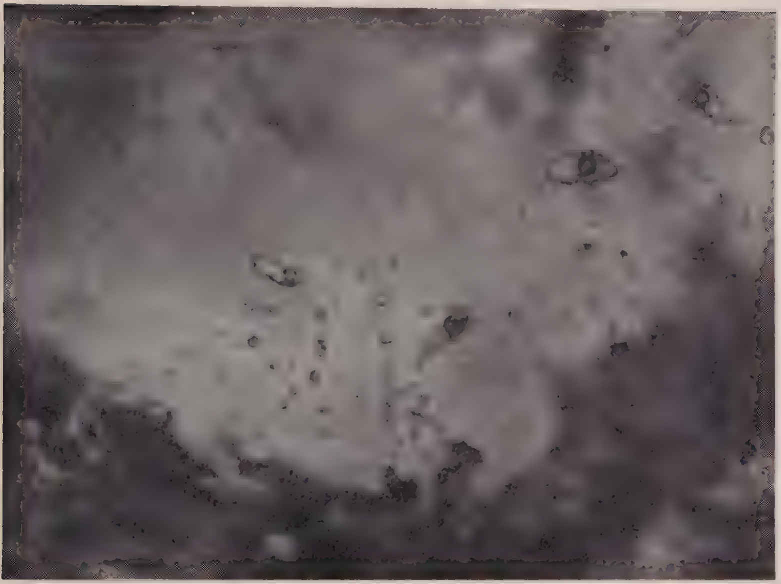

FIG. 4. Disseminated irregular inclusions in beryl, Strickland-Cramer pegmatite. ×200.

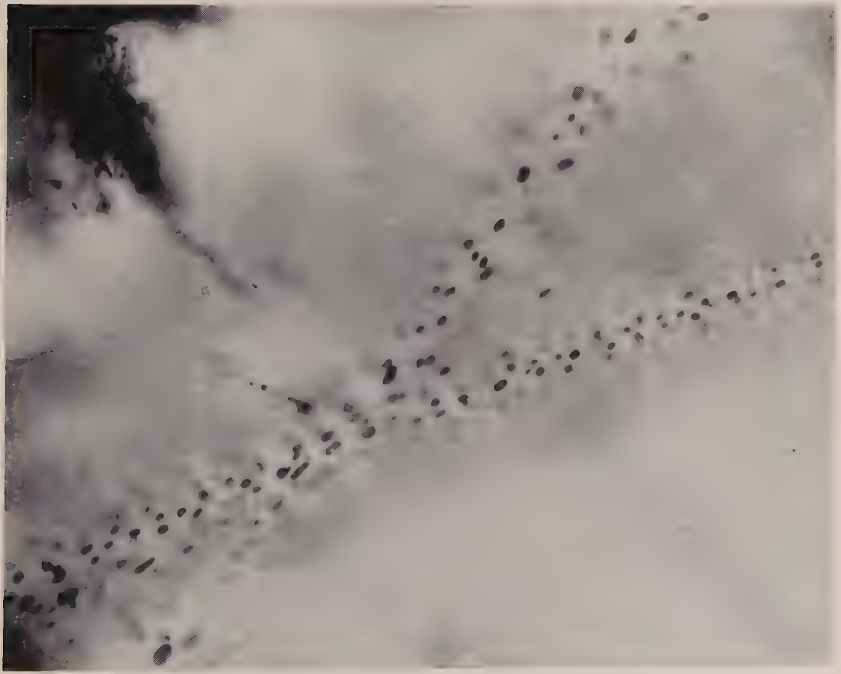

FIG. 5. Secondary inclusions along a branching healed fracture in beryl, Bordonaro north pegmatite. ×100.

Strickland-Cramer pegmatite is worth noting. This crystal is veined along its basal parting by quartz and cleavelandite deposited later in the development of the replacement body. The veins are several inches apart. Plates cut from this crystal show that the distribution of inclusions of type (1) in the beryl is unrelated to the veins. The inclusions occur in swarms away from the veins as well as adjacent to them. If the inclusions formed during or after the development of the veins, it seems extraordinary that they show no relation to so obvious a discontinuity.

As yet we find no absolute proof that inclusions of type (1) are primary, but it seems probable that material selected with specific reference to the problem of origin will show that inclusions of this type developed during beryl formation. For the present, but solely for convenient reference and distinction from the obviously secondary inclusions, we shall refer to them as the "apparently primary inclusions."

These inclusions are divisible into two main groups,* according to their behavior when heated. One group is represented by inclusions discovered in beryl from the Selden east pegmatite (Cameron, Rowe and Weis, 1951 p. 908) and subsequently recognized in beryl from the Strickland-Cramer pegmatite. At room temperature, each inclusion consists of a vapor phase plus a liquid phase. Heating causes the vapor phase to shrink and finally to disappear. Temperatures of disappearance (precise determinations only) are given in Fig. 6. For the Strickland-Cramer pegmatite, all but two of the precise temperature values shown in the figure fall within the range 210° C.–260° C. Approximate values for a large number of additional inclusions fall within the same range. For beryl from the Selden east pegmatite, values obtained range from 187° C. to 386° C.; most of the precise values plotted, together with approximate values for about 150 additional inclusions not shown in Fig. 6, are between 210° C. and 290° C. Nothing in the crystals themselves suggests an explanation of the range of values. Some inclusions less than a tenth of a millimeter apart give markedly different values of the temperature of disappearance. On the other hand, certain clusters of inclusions give values remarkably close to one another. The values obtained have no relation either to the sizes of inclusions or their positions within crystals.

The fluids composing inclusions of this type are of high mobility even at room temperatures, and their composition is of much interest. Fortunately, some of the tubular inclusions are highly elongate and of uniform diameter, hence the ratio of the length of the part of the cavity occupied by liquid at a given temperature to the total length of the

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* The grouping of inclusions in the present paper differs from that employed by the authors in a brief note in this journal, vol. 36, pp. 906–910. Additional work has shown that Group B recognized in that paper belongs under the second main group described below.

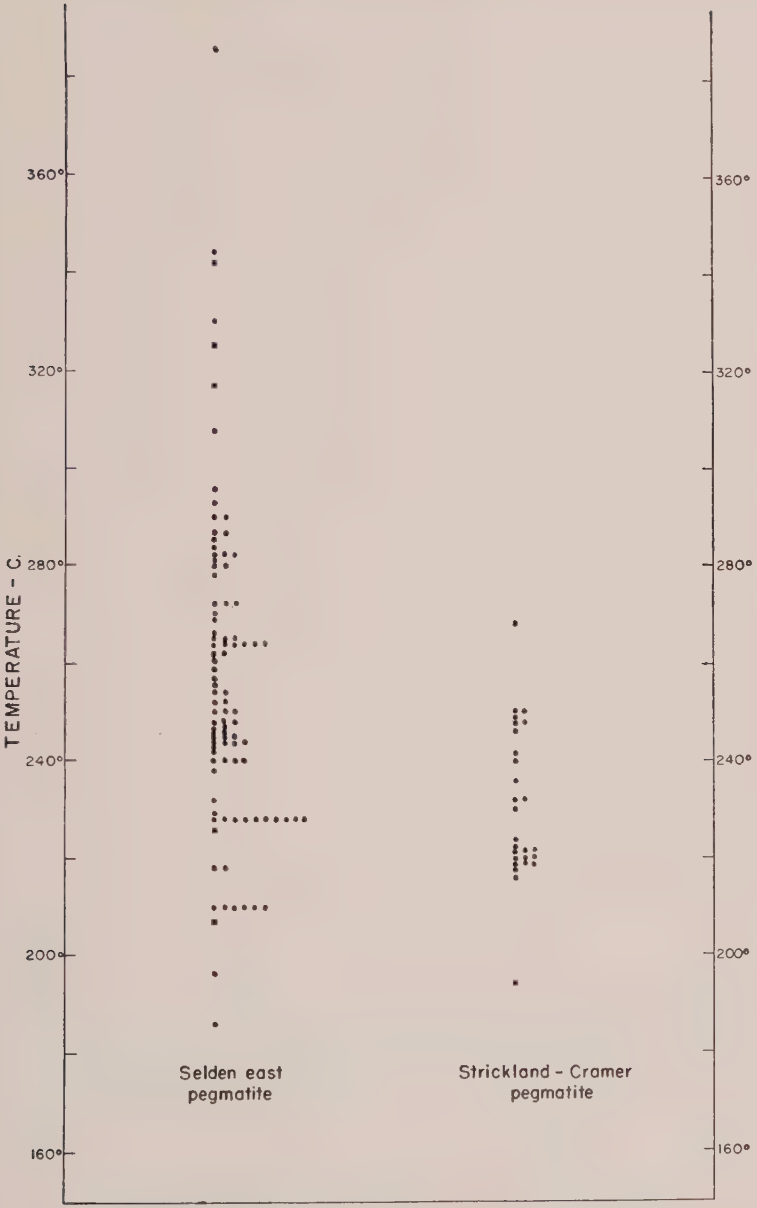

Fig. 6. Temperatures of disappearance of the vapor phase in two-phase apparently primary inclusions in beryl from the Selden east and Strickland-Cramer pegmatites. Each dot represents one inclusion.

cavity should closely approximate the ratio of volume of liquid to volume of cavity. Changes in this ratio with heating have been determined by Rowe for some of the tubular inclusions, and curves for a representative pair of these are shown in Fig. 7. With the curve for each inclusion a curve is also given for a hypothetical inclusion composed of pure water that would exactly fill its cavity at the temperature of filling of the actual inclusion. In both cases the two curves are very close to one another, and this suggests that the inclusions either approach pure water or are composed of some fluid remarkably similar to water in thermal characteristics. Furthermore, the melting points of the phase formed by freezing the liquid in 5 inclusions of this group in beryl from the Selden pegmatites range from −5° C. to −12° C. These data likewise suggest aqueous solutions. Inasmuch as the possible salts can only be guessed, the concentrations of the solutions can be estimated only roughly, but recorded data on melting points of solutions of various salts in water point to concentrations in the range between 2.7 and 6.5 molar.

The second group of apparently primary inclusions is considerably different from the first. At room temperature, these inclusions show three phases, -two liquid phases plus a vapor phase (Fig. 8, *A*, *C*) At about 31° C., the boundary between the vapor phase and one liquid phase fades and disappears, so that the inclusions consist only of the vapor phase plus a single liquid phase (Fig. 8, *B*, *D*). As 31° C. is the critical temperature of carbon dioxide, we infer that the disappearing phase is essentially liquid carbon dioxide. Owing to difficulties with our present cooling cell, we have not yet been able to determine the melting point of the phase.

All apparently primary inclusions observed in beryl from the Case No. 1 and Case No. 2 pegmatites, the Bordonaro north and south pegmatites, the Gotta-Walden pegmatite, and the Slocum pegmatite are of the three-phase type at room temperature. All those observed in beryl from the Case No. 2 pegmatite, and some of the inclusions in beryl from each of the other four pegmatites, show shrinkage and final disappearance of the vapor phase on heating. Temperature data (precise determinations only) for these inclusions are shown in Fig. 9. The ranges of values are as follows:

Pegmatite	*Number of inclusions*	*Temperatures of disappearance of the vapor phase*
Bordonaro north	23	282°–353° C.
Bordonaro south	2	299°–338° C.
Case No. 2	30	292°–332° C.
Slocum	4	310°–455° C.
Gotta-Walden	6	238°–355° C.

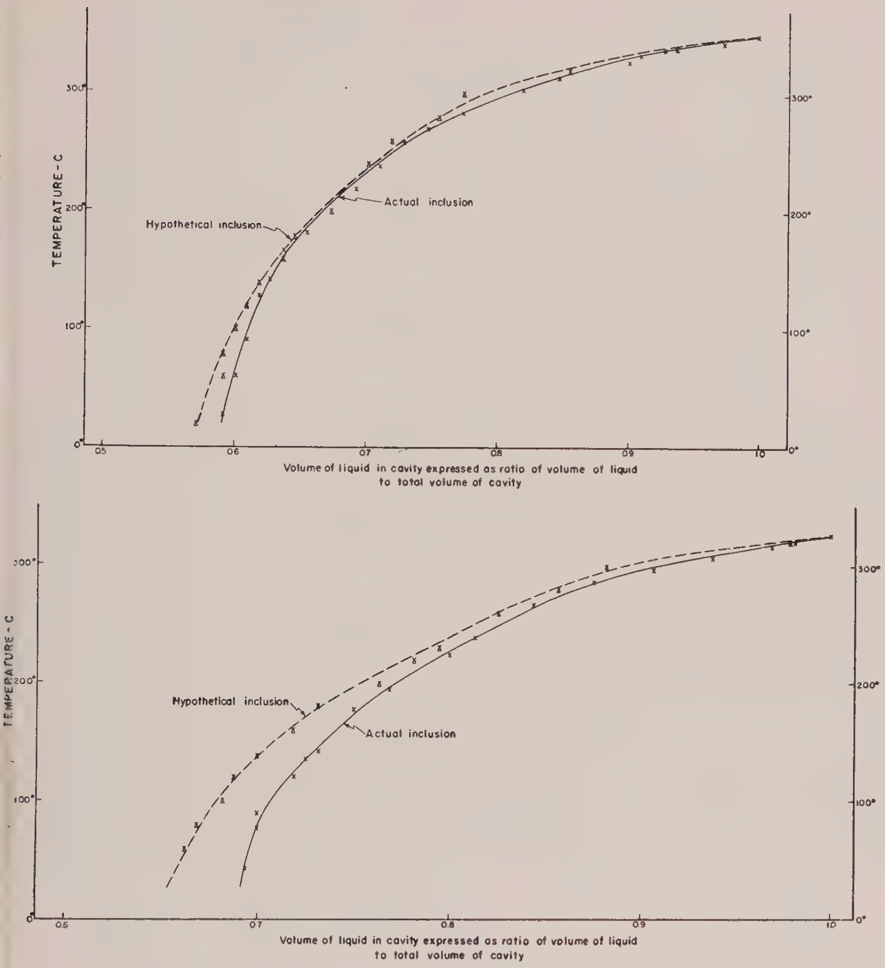

FIG. 7. Curves showing changes in volume for actual two-phase inclusions in beryl from Selden east pegmatite and for hypothetical inclusions composed of pure water. In each case, both the actual inclusion and hypothetical inclusion are filled with liquid at the temperature corresponding to the upper end of the pair of curves.

The changes in volume of liquid when an inclusion of this type is heated have been determined for 6 inclusions. In Fig. 10, the curves for two such inclusions are plotted against the comparable curves for a hypothetical inclusion composed of pure water that would be filled at the same temperatures as the actual inclusions. The curves for the three-phase inclusions differ markedly from those for the water inclusions, and furthermore lie on the opposite side of the curves for the water

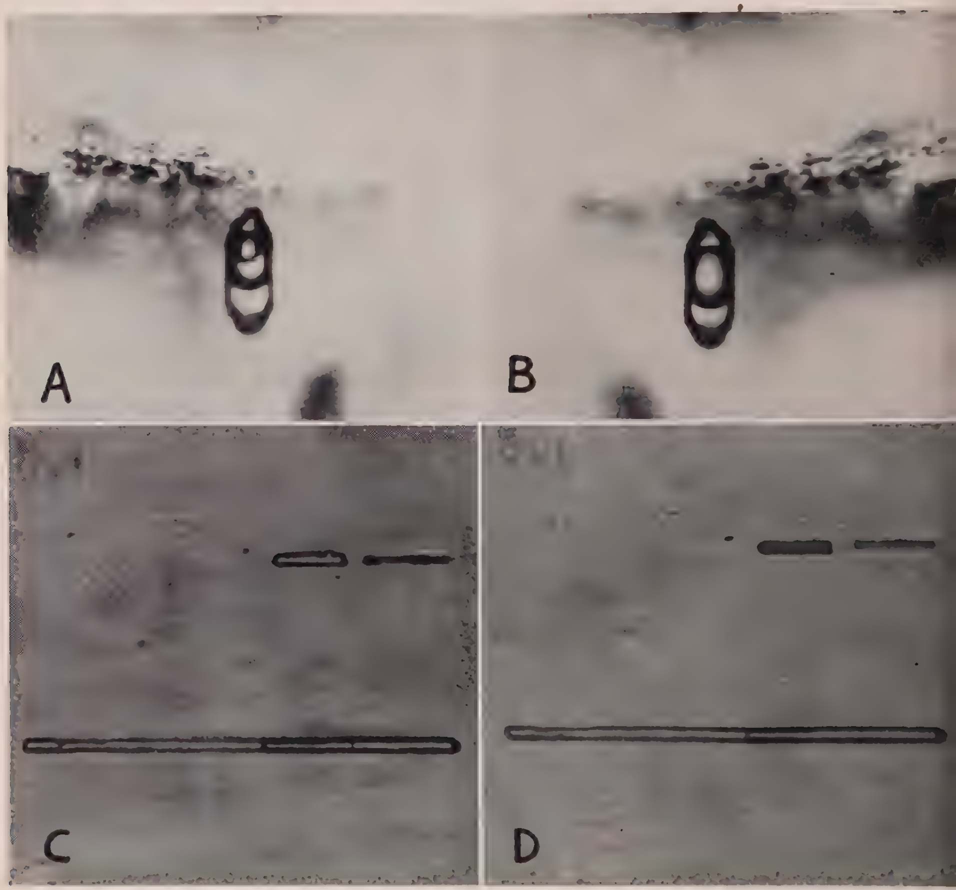

FIG. 8. Three-phase inclusions in beryl. (A) Oriented negative crystal inclusion in beryl, Bordonaro north pegmatite, at room temperature, showing 3 phases. ×200. (B) Same inclusion, heated abov 31° C. (C) Oriented tubular inclusion in beryl, Case No. 2 pegmatite, at room temperature, showin 3 phases. ×460. (D) Same inclusion, heated above 31°.

inclusion from the curves for the two-phase inclusion of Fig. 6. The presence of significant amounts of carbon dioxide in the three-phase inclusion is presumably responsible for the marked difference.

Many inclusions of the three-phase type, when heated, behave differently from those described. According to their precise behavior, they fall into three subgroups. In the first, the vapor phase expands until it occupies the entire cavity. Inclusions having this behavior occur in two beryl crystals from the Bordonaro north pegmatite, in one crystal from the Bordonaro south pegmatite, in one crystal from the Case No. 1

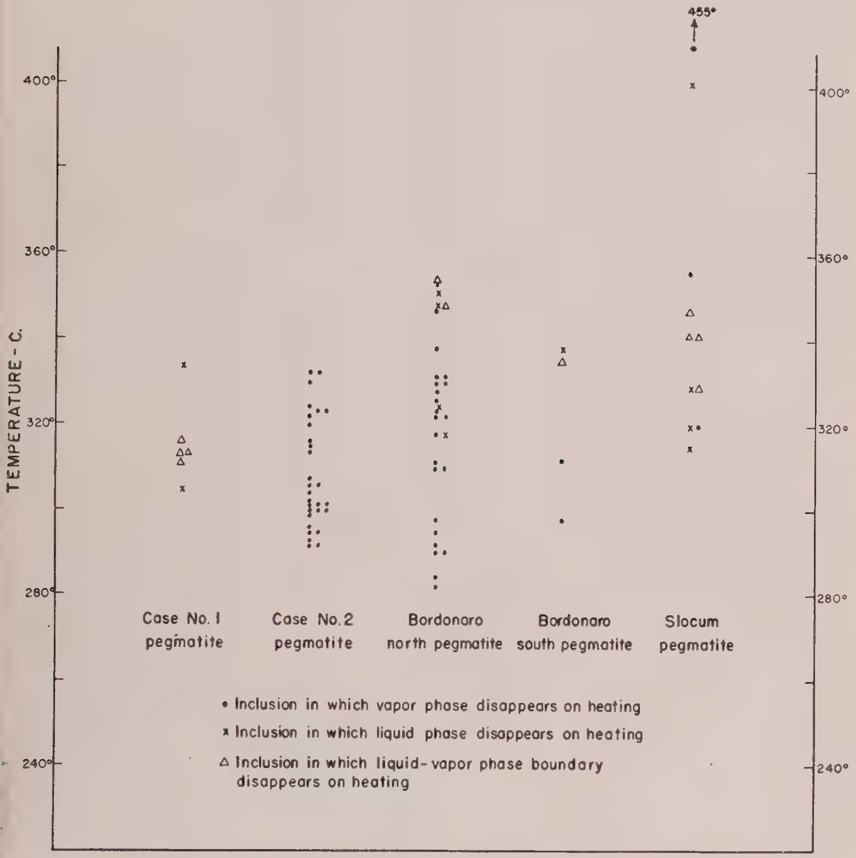

FIG. 9. Temperatures of disappearance of liquid phase, vapor phase, or liquid-vapor boundary in apparently primary three-phase inclusions in beryl from the Bordonaro north, Bordonaro south, Case No. 2, Slocum, and Case No. 1 pegmatites. Each symbol represents one inclusion.

pegmatite, and in two crystals each from the Slocum and Gotta-Walden pegmatites. Temperature data for these inclusions are also shown in Fig. 9, and may be summarized as follows:

Pegmatite	*Number of Inclusions*	*Temperatures of disappearance of the liquid phase*
Bordonaro north	3	324°–351° C.
Bordonaro south	1	338° C.
Case No. 1	2	305°–334° C.
Slocum	5	310°–401° C.
Gotta-Walden	15	310°–346° C.

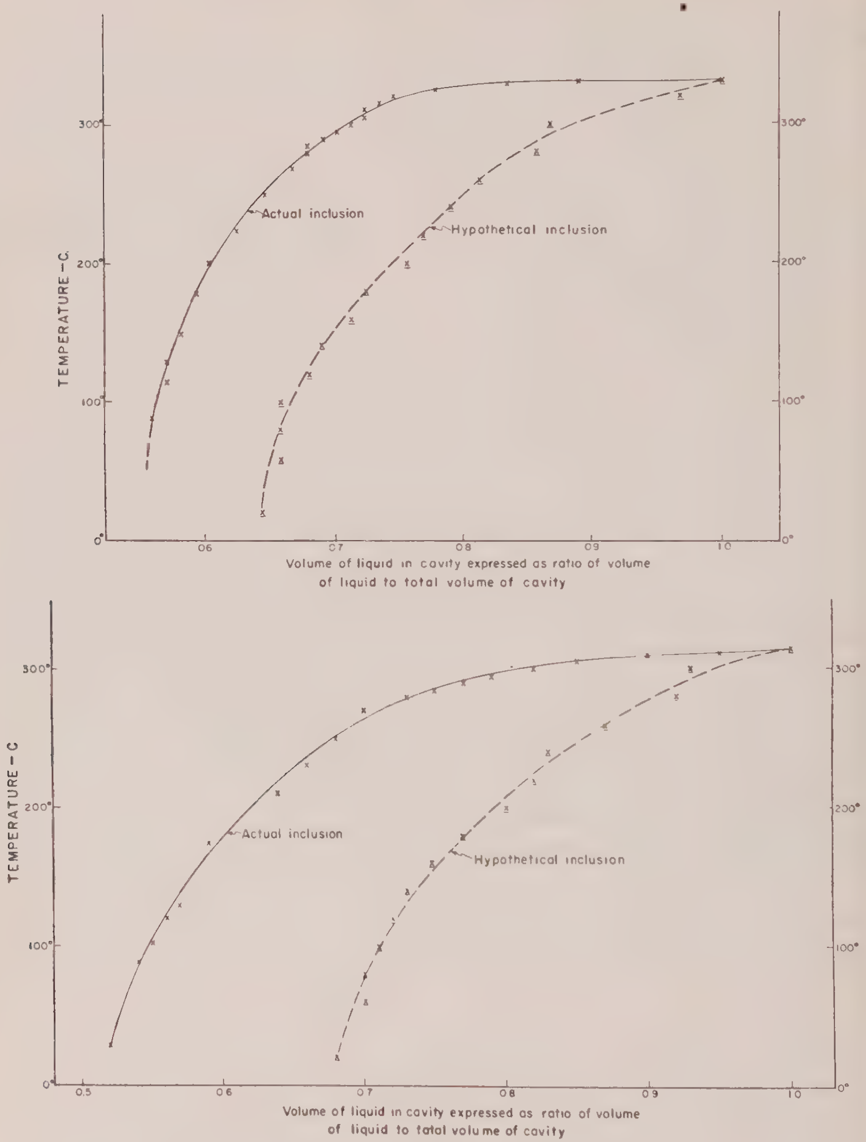

FIG. 10. Curves showing change in volume for actual inclusions of the three-phase type in beryl from the Case No. 2 pegmatite and for hypothetical inclusions composed of pure water. In each case, both the actual inclusion and the hypothetical inclusion are filled with liquid at the temperature corresponding to the upper end of the pair of curves.

In the second subgroup of inclusions, heating causes no change in liquid-vapor ratio, but the boundary between the two phases ultimately fades and disappears as though the critical temperature of the liquid phase were reached. Ingerson has shown (1947, p. 378, Fig. 1) that this behavior is to be expected if an inclusion consisting of water and vapor in the proper ratio (about 3:7) at room temperature is heated to the critical temperature of water. In the inclusions under discussion the system involved is not so simple, but the principle involved is presumably the same.

Inclusions of the second subgroup have been found in two crystals of beryl from the Bordonaro north pegmatite, one crystal from the Bordonaro south pegmatite, and one from the Slocum pegmatite. Temperature data for these inclusions are shown in Fig. 9 and may be summarized as follows:

Pegmatite	*Number of Inclusions*	*Temperatures of disappearance of the phase boundary*
Bordonaro north	3	348°–354° C.
Bordonaro south	1	335° C.
Slocum	4	329°–347° C.

In the third subgroup of three-phase inclusions, progressive heating beginning at room temperature first causes the presumed liquid carbon dioxide phase to disappear, then the vapor phase shrinks perceptibly, then the liquid-vapor boundary fades and disappears. One inclusion of this sort has been found in a crystal from the Bordonaro north pegmatite. Presumably further search would disclose others, hence a possible subgroup consisting of such inclusions must be recognized. In the lone example, the phase boundary disappears at 354° C. It seems probable that this type of inclusion is a variant of the second subgroup, having a slightly different density.

Few data for the disseminated rounded, elongate, or irregular inclusions of type (2) have been obtained. Temperatures of disappearance of the vapor phase in four two-phase inclusions of type (2) in pink beryl from the Gillette pegmatite range from 232° to 263° C. Ten similar inclusions in beryl from the Strickland-Cramer pegmatite gave values of 180° to 260° C. In a three-phase inclusion in beryl from the wall zone of the Bordonaro south pegmatite the liquid phase disappeared at 383° C. The data are too limited to justify conclusions, but it is of interest that the values are similar to those of comparable apparently primary inclusions.

Data for only 18 inclusions of Type (3) are at hand. Fourteen inclusions in beryl from the Case No. 1 pegmatite show disappearance of the liquid phase at 287° to 316° C. The vapor phase in one inclusion in beryl from the Bordonaro north pegmatite disappeared at 322° C., in another at 335° C. The vapor phase in one inclusion in beryl from the Slocum pegmatite disappeared at 310° C., in another at 336° C. The determinations were made early in the work, when the present procedure for determining the presence of a third phase in inclusions at room temperatures was not yet worked out. Subsequent work has shown that most secondary inclusions in beryl from these two pegmatites are three-phase inclusions.

Certain broad relationships between behavior groups of apparently primary inclusions in beryl and the occurrence of beryl within the pegmatites studied appear to be suggested by the data. The data for the Case No. 1, Case No. 2, Bordonaro, and Slocum pegmatites, are for inclusions in beryl occurring in the wall zones or intermediate zones of the pegmatites. Plagioclase, perthite, and quartz are the essential constituents of these zones, and the zones apparently represent the early and middle stages of crystallization of the pegmatites. Apparently the characteristic inclusions in beryl having this occurrence are three-phase inclusions containing abundant liquid carbon dioxide. The data for the Selden east and Strickland-Cramer pegmatites, on the other hand, are for beryl from the last-formed units of the pegmatites, a pod in the former case, replacement bodies in the latter case. Apparently the characteristic inclusions in beryl in these late units are free of detectible amounts of carbon dioxide.

The Gotta-Walden pegmatite may be an exception to this two-fold grouping, for three-phase inclusions occur in beryl from the apparent core. It is doubtful, however, whether the true core of this pegmatite has ever been seen, either in surface exposures or in drill cores. The mineral assemblage forming the apparent core is the same as that forming wall or intermediate zones in the Case, Bordonaro, and Slocum pegmatites.

Fluid inclusions in quartz: Although quartz proved unsatisfactory for determinations of temperature by the fluid inclusion method, intensive study of a large number of plates prepared from quartz from the Case No. 2 and Case No. 1 pegmatites has yielded a few data. Three types of inclusions have been found, as follows:

(1) Disseminated negative crystals.
(2) Disseminated irregular inclusions.
(3) Irregular to rounded inclusions arranged along cross-cutting planes or curved surfaces, many of which intersect or branch.

Inclusions of the first group are rare; those of the second group are not abundant. Inclusions of the third type are exceedingly abundant in every plate studied. In most of the plates cut, they are the only type of inclusion present. In the remainder, they generally outnumber the other types of inclusions. Inclusions of the first two groups may well be primary, especially the negative crystals; inclusions of the third group are obviously secondary.

Determinations have been made successfully for 17 inclusions of the negative crystal type in quartz from the Case No. 2 pegmatite. In 15 of the inclusions, one from the wall zone and the other 14 from quartz of the core, the vapor phase first shrinks slightly but perceptibly, then the phase boundary fades and disappears. Temperatures of disappearance range from 293–357° C. In the other two inclusions, in a plate of quartz from the core, the vapor phase shrinks and finally disappears, at 352° C. in one inclusion, at 395° C. in the other.

Temperatures of disappearance of the vapor phase in 29 disseminated irregular inclusions in quartz from the core of the Case No. 2 pegmatite range from 124° C. to 292° C., but values for 27 of the inclusions range from 193° C. to 292° C. In 6 other inclusions from quartz of the core, and in one in quartz from the wall zone, the liquid-vapor boundary fades and disappears, at temperatures ranging from 320° C. to 390° C.

Four secondary inclusions of type (3) in quartz from the Case No. 1 pegmatite gave temperatures of disappearance of the vapor phase ranging from 316°–341° C. All other secondary inclusions of this type studied leaked at temperatures at which the liquid phase fell considerably short of filling the cavities.

Most of the inclusions of each of the groups described above are now known to consist of three phases at room temperature, like the inclusions in the associated beryl. A very few appear to consist of two phases. The remainder are so small that the presence or absence of a third phase can not be established.

DISCUSSION OF RESULTS

Assumptions involved in the use of fluid inclusions:

The use of fluid inclusions rests on a series of assumptions (Ingerson, 1947, pp. 376–377; Bailey and Cameron, 1951, p. 642). These must be reviewed with each new investigation of fluid inclusions, so that their validity with respect to the inclusions studied can be appraised. Review is particularly necessary for pegmatite minerals, because the physicochemical conditions under which pegmatites crystallize are uncertain. The assumptions ordinarily made are as follows:

(1) The cavity occupied by the inclusion was completely filled with liquid at the time of crystallization.
(2) Primary fluid inclusions can be distinguished from secondary inclusions.
(3) The liquid is an aqueous solution containing no carbon dioxide or other gas in large concentration.

(4) The pressure on the liquid at the time of inclusion was small, or its magnitude can be estimated.
(5) There has been no significant change in the volume of the cavity itself due to pressure, solution, or precipitation.
(6) There has been no addition of liquid to the cavity or loss of liquid from it.
(7) Representative samples are used for temperature determinations.

For liquid inclusions in pegmatite minerals, this is a formidable list of assumptions, quite apart from the results of the present investigation. Even the seventh cannot be accepted as justified for the beryl studied. The sections of crystals cut give a generous sampling of beryl in the pegmatites involved, except for the Gillette, Case No. 1, and Slocum pegmatites. Beryl studied from the Strickland-Cramer pegmatite is satisfactorily representative of beryl in the replacement bodies, though not of beryl in other units. However, many of the mineral sections cut contain no usable inclusions, and additional sections must be made before we can be sure that representative sampling has been attained.

The other assumptions are difficult matters. The first assumption can ordinarily be made with considerable assurance for inclusions in minerals that are known to have formed from solutions at low temperatures. This is not so for inclusions in pegmatite minerals. The physico-chemical nature of the fluids from which pegmatite minerals crystallize is decidedly uncertain, and there is no *a priori* assurance that the cavities in beryl were filled with liquid at the time the crystals formed. Depending upon one's ideas of the physical chemistry of pegmatitic fluids, the mother fluids of pegmatites may be regarded as consisting of a liquid phase below its critical temperature, a fluid phase above its critical temperature, a combination of a liquid phase plus a vapor phase (Bowen, 1933, pp. 115 118) or even (Smith, 1948*b*; Neumann, 1948) two immiscible liquid phases. Furthermore, it is conceivable that the number and kinds of phases (that is, whether vapor or liquid) present in the fluid may not be the same in the early stages of crystallization of pegmatites as in the later stages. The phases present will depend on relative concentrations of the various constituents, the hyperfusibles in particular, on the pressure within the system, and on the temperature.

How shall one distinguish among fluid inclusions produced under these various sorts of conditions? Let us assume that the pegmatite fluid is a liquid and is at a temperature below its critical temperature. Inclusions formed in crystals growing in such a liquid should fill their cavities at the instant of entrapment; the liquid phase should shrink and a vapor phase should appear and grow as the crystal cools to room temperature. If now the crystal is heated, the liquid phase should expand and ultimately fill the cavity. This would be, then, an inclusion of the type normally envisioned in connection with use of the Sorby method. Among the inclusions here described, those in beryl from the Selden east, Strickland-Cramer, and Gillette pegmatites would, *so far as their behavior indicates*, appear to qualify as examples.

Let us suppose, on the other hand, that pegmatites form from fluids above their critical temperatures. The fluid entrapped will again consist of a single phase that will completely fill the cavity. Its behavior on cooling, however, will depend on its original density, which is a function of pressure. If its density exceeds the critical density, then as the crystal cools below the critical temperature, the inclusion will pass directly into a liquid phase that will completely fill the cavity. No second phase will appear until the cooling liquid shrinks so that it can no longer fill the cavity. At this point, a vapor phase will appear. When beryl containing such an inclusion is reheated, its behavior will be indistinguishable from that of an inclusion formed from a liquid below its critical temperature. Thus, using the data given by Kennedy (1950, p. 540, Fig. 3) both an inclusion formed of water at a temperature of 360° C. and at a pressure of about 440 bars and one formed at 555° C. and a pressure of 2120 bars would show disappearance of the vapor phase at 325° C. If we assume that the inclusions of the Selden east pegmatite are 5-molar aqueous solutions, then from the curve given by Ingerson (1947) for solutions of KCl and NaCl in water, the critical temperature of the inclusions would probably be greater than 400° C. In the absence of corroborative data, we cannot beg the question by assuming that pegmatite liquids do or do not attain such temperatures. The one conclusion possible is that regardless of whether inclusions of the type found in the Selden east pegmatite were entrapped at temperatures above or below their critical temperatures, the temperatures of disappearance of the vapor phase recorded for these inclusions are less than the true temperatures of crystal formation, provided the other assumptions underlying the Sorby method hold for these inclusions. If properly corrected for pressure, however, the temperatures of disappearance should give the temperatures of formation of the inclusions.

If the fluid trapped in an inclusion is above its critical temperature, but has a density less than the critical density, then three cases are possible. If the density is only slightly less than the critical density, then as the inclusion is cooled below the critical temperature, a liquid phase will appear. The liquid will not fill the cavity, however, hence a vapor phase will be present. As the crystal is cooled to room temperature, the liquid phase will shrink. This is the first case. The amount of shrinkage will decrease with decreasing density of the inclusion at the time of entrapment, until a certain value of the density is reached for which the shrinkage of the liquid phase due to cooling is compensated by condensation of the vapor phase. This is the second case. In both cases, the inclusion at room temperature will consist of a liquid phase and a vapor phase. In the first case, the liquid phase will expand as the inclusion is heated, but before it expands sufficiently to fill the cavity, the liquid-vapor boundary will fade and disappear. In the second case, the liquid phase will neither shrink nor expand, but when the critical temperature of the liquid is reached, the liquid-vapor boundary will fade and disappear. In both cases, the temperatures of boundary disappearance will be less than the temperatures of crystal formation. The third case occurs when the density of the liquid at the time of entrapment falls sufficiently below that of the second case. In this third case, a liquid phase will appear and will expand as the inclusion is cooled to room temperature, and conversely, heating will cause the liquid phase to shrink and disappear at a temperature below the temperature of crystal formation.

If a fluid inclusion shows disappearance of the liquid phase or of the vapor phase, and the pressure is known, then appropriate correction for the pressure will give the temperature of formation of the inclusion, provided the inclusion satisfies the other assumptions of the Sorby method.

Finally, let us suppose that a beryl crystal happens to be forming at a time when minute bubbles of vapor are rising through a liquid phase. In such an environment, two kinds of inclusions may conceivably be trapped within a crystal, one type consisting of liquid only, the other consisting of liquid plus a bubble of vapor. Each will consist of a liquid phase and a vapor phase at room temperature. The first type, when heated, will show shrinkage and final disappearance of the vapor phase. If the inclusion satisfies the other assumptions of the Sorby method, the temperature of filling, corrected properly for pressure, will be the correct temperature of formation of the inclusion.

The behavior of the second type will depend on the ratio of liquid to vapor at the time of entrapment. For a certain critical ratio, the result will be an inclusion in which the liquid-vapor ratio is unchanged with cooling. When such an inclusion is heated, the phase boundary will not shift, but will fade and disappear when the critical temperature of the liquid is reached. The temperature of disappearance gives the critical temperature of the fluid composing the inclusion but gives no information as to the temperature at which the inclusion formed. If the liquid-vapor ratio at the time of entrapment exceeds the critical ratio, the liquid will expand until it fills the cavity, but the temperature of filling, corrected for pressure, will be higher than the temperature of formation of the inclusion. If the liquid-vapor ratio is lower than the critical ratio, the vapor phase will expand as the inclusion is heated and will ultimately fill the cavity. The temperature of filling, corrected for pressure, will be either lower or higher than the true temperature of formation of the crystal, depending on the exact ratio (see Ingerson, 1947, Fig. 1 and p. 386).

If inclusions in a crystal have formed under such conditions, it seems probable that primary inclusions of all three types described above will occur together. If crystal after crystal within a given unit of a pegmatite shows all three or even two of these types, a two-phase mother liquor is a definite possibility. However, the same feature can also result from differential leakage of inclusions or from differential addition of material to inclusions.

The conditions outlined above, the possible types of inclusions formed, their appearance at room temperature, and their behavior when heated are summarized in Table 2. In the final two columns is given the occurrence of possible examples among the apparently primary inclusions in beryl that we have studied. Inclusions that consist of two phases at room temperature and those that consist of three phases are both included because for purposes of the foregoing analysis the composition of the inclusions makes no difference, and the existence of a third phase at low temperatures is extraneous to the discussion.

Several significant points have been brought out by compilation of the table. One is that all the types of apparently primary inclusions in beryl (and in quartz also) can be accounted for in terms of the possible phase conditions defined in the table. A second is that differences in degrees of filling and behavior of inclusions within a single crystal can be explained in terms of Case IV of the table without need of variations in either temperature or pressure within the period of formation of the crystal, and without need of differential leakage or differential addition. Explanation in terms of Case I, Case II or Case III, however, requires significant variations in temperature or pressure, or both, during the period of crystal formation, or requires differential leakage from, or addition to, the inclusions after their formation. A third conclusion is that nothing in the nature of a fluid inclusion at room temperature or in its behavior when heated tells us whether inclusions in which the liquid phase expands and fills the cavity belong under Case I, Case II, or Case IV; i.e., whether they have formed under supercritical or subcritical conditions

TABLE 2. KINDS AND BEHAVIORS OF POSSIBLE INCLUSIONS IN PEGMATITE MINERALS

Nature of medium at time of crystal formation			Nature of possible fluid inclusions at time of entrapment			Nature of possible fluid inclusions at room temperature	
			Liquid	Gas	Liquid plus vapor	Liquid plus vapor	Liquid (1) plus liquid (2) plus vapor
I. Fluid, below critical temperature; no vapor phase present			x			x	
			x				x
II. Fluid, above critical temperature; density greater than critical density				x		x	
				x			x
III. Fluid, above critical temperature; density less than critical density	(1) Density near critical density	(a)		x		x	
		(b)		x		x	
	(2) Density moderately lower than critical density	(a)		x			x
		(b)		x			x
	(3) Density much below critical density	(a)		x		x	
		(b)		x			x
IV. Liquid plus vapor	(1) Liquid only		x See under Case I				
	(2) Liquid-vapor ratio markedly higher than critical ratio				x (a)	x	
					(b)		x
	(3) Liquid-vapor ratio equal to critical ratio				x (a)	x	
					(b)		x
	(4) Liquid-vapor ratio less than critical ratio				x (a)	x	
					(b)		x

TABLE 2. KINDS AND BEHAVIORS OF POSSIBLE INCLUSIONS IN PEGMATITE MINERALS (*continued*)

Behavior when heated					Occurrence of possible examples among inclusions studied		
Vapor phase disappears	Liquid (2) disappears then vapor phase disappears	Liquid phase(s) disappear(s)	Liquid (2) disappears (*) Liquid (1) expands somewhat, then liquid-vapor boundary fades and disappears	Liquid (2) disappears (*) Liquid (1) vapor boundary fades and disappears	Pegmatite	Mineral containing the inclusions (†)	
x					Selden east Strickland-Cramer Gillette Case #2	Beryl Beryl Beryl Quartz	*D*
	x				Bordonaro north Bordonaro south Case #2 Slocum Gotta-Walden	Beryl Beryl Beryl Beryl Beryl	*A*
x					Selden east Strickland-Cramer Gillette Case #2	Beryl Beryl Beryl Quartz	*D*
	x				Bordonaro north Bordonaro south Case #2 Slocum Gotta-Walden	Beryl Beryl Beryl Beryl Beryl	*A*
			x		None		
			x		Bordonaro north	Beryl	
				x	None		
				x	Bordonaro north Bordonaro south Case #1 Slocum	Beryl Beryl Beryl Beryl	*B*
		x			None		
		x			Bordonaro north Bordonaro south Gotta-Walden Case #1 Slocum	Beryl Beryl Beryl Beryl Beryl	*C*
x					Selden east Strickland-Cramer Gillette Case #2	Beryl Beryl Beryl Quartz	*D*
	x				Bordonaro north Bordonaro south Case #2 Slocum Gotta-Walden	Beryl Beryl Beryl Beryl Beryl	*A*
				x	None		
				x	Bordonaro north Bordonaro south Case #1 Slocum	Beryl Beryl Beryl Beryl	*B*
		x			None		
		x			Bordonaro north Bordonaro south Gotta-Walden Case #1 Slocum	Beryl Beryl Beryl Beryl Beryl	*C*

(*) By these two columns, both inclusions consisting of two phases at room temperature and inclusions consisting of three phases at room temperature are covered. Heating of the three-phase inclusions first causes disappearance of one liquid phase, after which the inclusions behave as indicated by the headings of the columns.

(†) Each letter designates a particular group of inclusions. Thus the inclusions of group A cited as examples of Case I are the same as those of group A cited as examples of Cases II and IV (1) *a*.

from single-phase or two-phase systems. This fact is implied in part in discussions by Ingerson (1947), Scott (1948), Smith (1949), and Kennedy (1950) but has not been specifically stated.

A final conclusion is that even if one accepts the definitions of "hydrothermal" and "pneumatolytic" fluids proposed by Smith and his co-workers (1950, p. 587), the statement (p. 584) that

> "If the liquid phase fills the inclusions before the vapor phase during heating, the transporting agent was a hydrothermal solution. Conversely, if the vapor phase fills the inclusion before the liquid phase during heating the transporting agent was a pneumatolytic mixture."

holds only if it can be shown that the system does not fall under Case IV of the table, and that there has been no leakage or addition.

Table 2 indicates that even if inclusions can be shown to satisfy other assumptions underlying the Sorby method, it is by no means easy to interpret temperature data with reference to the first assumption. When the validity of the other assumptions is appraised, difficulties multiply at a disconcerting rate.

The second assumption, that primary inclusions can be distinguished from secondary inclusions, is far less easy to satisfy than a perusal of the literature would indicate. Criteria for distinguishing primary from secondary inclusions have been discussed by Laemmlein (1929). He gives the following as characteristics of primary fluid inclusions:

(1) They are always associated with real crystal surfaces.
(2) They may be arranged parallel to growth planes of a crystal, or along a surface that records the direction of growth of an edge of a crystal; or along a row or line that traces the growth direction of the apex of a crystal.
(3) They may consequently show a zonal arrangement corresponding to the growth zones of a crystal.
(4) When arranged along planes, these planes do not intersect.

By contrast, he found experimentally that whereas some secondary inclusions formed by healing of fractures are negative crystals, others are rounded, oval or irregular. He pointed out that fluid inclusions formed by partial healing of fractures occurred along planes that cut across growth zones and other primary features of crystals and may branch or intersect. Laemmlein's discussion does not take fully into account the formation of primary inclusions along lineages.

In his discussion of secondary inclusions, Laemmlein appears to have considered only their development by the partial healing of cracks in crystals. Little consideration has been given to the possibility that fluid inclusions, negative crystals, or rounded or irregular inclusions may be of secondary origin even if they are disseminated through a crystal and appear unrelated to cracks or other features that might offer channelways for later solutions or might influence selective solution. Ingerson (1947, p. 377) concluded that one is probably warranted in assuming that inclusions in crystals that line vugs and other open spaces in veins are primary. He regarded inclusions along planes that cross grain boundaries without offset or change in orientation as almost certainly secondary, and stated that if planes of inclusions have a more or less constant orientation or orientations over an entire thin section, the chances are good that the inclusions are secondary, even those in planes not observed to cross grain boundaries. He further stated that where planes of inclusions are confined to individual mineral grains or where arrangement in planes is not evident, the chances are better that the inclusions are primary. He also pointed out that though control of orientation by the crystal lattice of grains can be evidence that inclusions are primary, secondary inclusions may also be thus controlled. Disseminated inclusions, related or unrelated to crystallographic elements of the host crystal, are not discussed in his article, nor is the question of irregular inclusions versus negative crystals discussed. He figures, however, two irregular inclusions in quartz that are presumably of the disseminated type. Twenhofel (1947), based his determinations of temperature on inclusions that apparently are arranged along growth zones of a crystal. There seems little doubt that the inclusions he studied were primary. Those who have developed and employed the decrepitation method (Scott, 1948; Peach, 1949; Smith, 1948*a*; Peach, 1951) have recognized the importance of distinguishing primary from secondary inclusions, but have discussed the means of distinguishing the two only briefly; indeed, the difficulty of interpreting the results of decrepitation in terms of primary and secondary inclusions appears to the writers to be a basic weakness of the decrepitation technique, and our results for quartz from the core of the Case No. 2 pegmatite confirm this view in one instance. Peach (1951, p. 34) states that secondary inclusions in pegmatite minerals were recognized by the relatively small volume of the bubbles and by their tendency to occur in planes within the mineral. Primary inclusions were recognized by their random distribution and lower degree of filling. Are these acceptable criteria? The writers would be reluctant to accept the highly irregular inclusion in quartz figured by Peach (1951, Plate 1, *A*) as primary without corroborative evidence. We have found secondary inclusions in quartz and other minerals that are remarkably similar. Buerger concluded from a study of lineages in crystals (1932*a*, 1932*b*), however, that primary cavities of various shapes, some highly irregular, are common along lineage structures in crystals.

A perusal of the literature suggests that the "apparently primary" inclusions in beryl and quartz discussed in the present paper would be accepted as primary without question by at least the majority of investigators, and that the disseminated irregular inclusions would likewise be accepted as primary. Furthermore, we find no record of experiments showing that such inclusions can develop by secondary processes. Nevertheless, we feel that concrete proof of primary origin of the inclusions is necessary before conclusions based on them can be fully accepted. The student of ore minerals knows all too well that disseminated solid inclusions can be developed in ore minerals by replacement, and selective sericitization and other alterations in zoned feldspars are likewise familiar to the petrographer. Perhaps liquid inclusions can develop in a similar manner.

The best evidence for primary origin of fluid inclusions appears to be as follows:

(1) Inclusions distributed or oriented with reference to growth zones or other primary features of crystals and showing no relation to cracks or other secondary features.
(2) Marked contrast in occurrence and in liquid-vapor ratios to clearly secondary inclusions in the same crystals.
(3) A systematic pattern of temperature variation from part to part of a crystal, or from mineral to mineral within a deposit, that is consistent with paragenetic sequence and with the geologic characteristics of a deposit.
(4) Clear cut evidence, in the form of cross-cutting relationships, that the inclusions were present before the formation of later pegmatite minerals; e.g., transection of inclusions by the walls of veinlets formed by later minerals.
(5) Agreement between temperatures of formation, as determined from the inclusions, and other geologic evidence.

If a group of fluid inclusions satisfies *both* (1) and (2) above, the probability that they are of primary origin is high. If a certain type of inclusion is present in crystal after crystal from a given deposit, and shows a consistent contrast in characteristics to inclusions related to secondary features, the chances that they are secondary are slight (Bailey and Cameron, 1951, p. 643). The case for primary origin becomes even stronger if the results of investigation satisfy conditions (3) and (5). It must be recognized, however, that only condition (4) constitutes unequivocal proof of primary origin. So far as the writers know, no case of this kind has been reported in the literature.

From the foregoing considerations, it seems necessary to conclude that primary and secondary inclusions are difficult to distinguish even in transparent minerals, in which the characteristics and distribution of inclusions can be studied. Such study is impossible for opaque minerals, and much though we admire the ingenuity of the decrepitation technique, this is a basic weakness of such fundamental importance that it makes all temperature data based on decrepitation of opaque minerals suspect of serious error.

The third and fourth assumptions can be considered together. Temperature data obtained by study of liquid inclusions require correction for the pressure at time of crystal formation. As Kennedy (1950, p. 543) has pointed out and others have recognized, this correction is based on the assumption that the coefficient of thermal expansion and the compressibility of the liquid in a fluid inclusion are essentially the same as those of pure water. This assumption is commonly forced on investigators because the compositions of fluid inclusions are little known and because temperature-pressure-volume data for solutions of various compositions at high temperatures and pressures are not available. It is generally assumed that fluid inclusions are dilute aqueous solutions, that the compressibilities of these solutions are not far different from those of pure water, and that the pressure corrections read from various charts showing P-V-T relationships for pure water (Ingerson, 1947; Scott, 1948; Kennedy, 1950) are of the right order of magnitude. This piling of assumption on assumption is certainly one of the weaknesses of the Sorby method. The errors thus introduced cannot be evaluated accurately at present. For deposits formed at shallow depths, such as the lead-zinc deposits of the Upper Mississippi Valley (Bailey and Cameron, 1951), pressure corrections required must be small and the errors introduced not serious. Both Ingerson (1947, p. 386) and Kennedy (1950, pp. 542–543) point out that for the deeper-seated deposits, however, errors of 50 per cent or more may be involved. If a gas of low critical temperature, such as CO_2, is present, no pressure correction can be calculated, because data for systems consisting of H_2O plus CO_2, H_2S, and other gases at high temperatures are lacking.

Even if an inclusion is known to consist solely of water, the usual pressure correction based on the P-V-T tables for water can be applied only if we know which of the four cases of Table 2 is represented. This may or may not be possible. Inclusions in minerals from deposits regarded on geologic evidence as formed at low temperatures generally show filling of the cavity by the liquid phase when the inclusions are heated, and it is probably safe to assume that these fall under Case I. The deeper-seated the conditions under which a deposit has formed, however, the greater is the uncertainty. For pegmatites, as we have seen, the uncertainty is very great.

From the standpoint of pressure correction, it makes no difference whether Case I or Case II is involved. The problem is to recognize whether either is involved. If all inclusions in a crystal or growth zone of a crystal show disappearance of the vapor phase on heating and show the same liquid-vapor ratio, the presumption is strong that they fall under Case I or Case II, although it is still possible that the inclusions have undergone uniform change in content (see below).

Similarly, if all the inclusions of a crystal or a growth zone of a crystal show *one* of the types of behavior described under Case III of the table, the presumption is strong that the inclusions fall under this case. Pressure corrections for this case are not given by ordinary curves, but can be calculated from the liquid-vapor ratios at room temperature if the pressure at time of formation is known.

Inclusions falling under Case IV, as discussed previously, should show a variety of behavior. A pressure correction can be applied only to those inclusions that consisted wholly of liquid at the time of entrapment. These would be the inclusions showing disappearance of the vapor phase at the lowest temperature.

There is still the problem of estimating pressure. As pointed out by Ingerson (1947, p. 386) and Kennedy (1950, p. 542), we rarely can determine whether this should be taken as the hydrostatic pressure, the pressure due to the weight of overlying rocks, or local pressures. Pressure at time of crystal formation is generally taken as equivalent to the weight of overlying rocks (e.g., Ingerson, 1947; Peach, 1951), but it is recognized that this is a statistical view of the problem. Smith and his associates (Smith, 1948*a*; Peach, 1951) rely on the pyrite geothermometer in calculating pressure corrections to be applied to temperature values obtained by decrepitation. The method is subject to whatever errors may be inherent in the technique, plus possible serious errors arising from the fact that neither the nature nor the origin of the inclusions presumed to be responsible for decrepitation in pyrite can be determined.

For the inclusions in beryl and quartz discussed in the present paper, no clear-cut estimate of pressure at time of formation can be made. If the pressure be taken as due to the weight of the overlying rock, then a pressure corresponding to a depth of 15,000 feet is probably to be regarded as a minimum. We must recognize, however, that the actual pressure may be unrelated to thickness of rock cover.

The fifth assumption underlying the Sorby method concerns possible changes in the volume of the cavity itself due to pressure, to solution of the walls, or to precipitation of solute upon the walls. This subject has been discussed by Ingerson (1946, pp. 376-377) and by Bailey and Cameron (1951, pp. 644-645). It seems probable that errors introduced in this way are minor.

The final assumption to be discussed is that involving addition or loss of fluid or vapor since the inclusions were formed. This is a question that has been given emphasis by Kennedy's discussion (1950), which is backed by experimental evidence that the fluid content of inclusions in fluorite can be increased by immersion in water under high pressure. Increase in fluid content lowers the temperature of disappearance of the vapor phase, whereas decrease in fluid content increases the temperature. It is not necessary, moreover, that there be change in quantity of fluid but only a change in composition, for the compressibility and coefficient of thermal expansion of the fluid will thereby be influenced even if the quantity of fluid remains unchanged.

It is obvious that no definite answer can be given to this question. Kennedy reasons that in veins the fall in temperature of the enclosing rocks is probably more rapid than the fall in pressure, and that the tendency will be for migration of material into fluid inclusions, so that temperatures of disappearance of the vapor phase will be lowered. Whether this is the case for pegmatites is largely a matter of opinion; we have no direct knowledge of relative rates of decrease of temperature and pressure after the formation of pegmatites. Furthermore, we know little of the pressure gradients required to induce migration of material or of the rates of migration. The pressure gradient of 1500 bars employed in the experiments cited by Kennedy was extreme. Kennedy's point, however, is one that merits consideration.

Smith and his co-workers (Smith *et al.*, 1950, p. 584) have pointed out that if adjustment of the contents of fluid inclusions to changes in pressure environment is an effective process, then it is difficult to explain why quartz from gold deposits of the Canadian shield contains two sets of inclusions exhibiting different degrees of filling. It might therefore be argued that because primary and secondary inclusions in quartz from the Case No. 2 pegmatite are similar in degrees of filling, the contents of the primary inclusions have been changed. Field study of the pegmatite suggests that the secondary inclusions could readily have formed at a later stage in the development of the pegmatite, however, and it is entirely possible that they developed within the same temperature-pressure range.

Two features of the inclusions described in the present study, however, require mention. For the carbon dioxide-bearing inclusions the pressure gradient is at present the reverse of that envisioned by Kennedy, for the pressure within the inclusions at room temperature must be close to the critical pressure (73 atm.) of carbon dioxide.

The second feature is the occurrence of apparently primary inclusions of different degrees of filling within a single crystal, in places within a cubic millimeter. This is difficult to explain if material has migrated in response to an external pressure gradient, unless there are complicating factors.

Further than the above, the effects of migration of material are difficult to appraise. It may well be that the spread of temperature values shown by inclusions in certain crystals or by various crystals from a single pegmatite is due partly or wholly to this cause.

A final source of error is one pointed out by Dreyer, Garrels, and Howland (1949). Owing to exothermic reactions and poor conduction of heat away from the surface of a growing crystal, the temperature of a liquid inclusion may exceed that of the surrounding liquid. This is a difficult problem for the student of pegmatites, because neither the reactions involved in crystallization nor the rates of heat conduction are known. However, crystallization from a solution is itself an exothermic reaction, hence heating of the liquid immediately adjacent to the face of the growing crystals is a definite possibility.

Interpretation of the Data

General Statement: It is evident that there are numerous possible sources of error in interpreting temperature data for pegmatite minerals. If the oriented tubular inclusions and negative crystals are not primary, the temperature values obtained are not related to temperatures of crystal formation, but to temperature conditions during some later period in the history of the pegmatites. The cavities in some of the inclusions may not have been filled with liquid at the time they formed. No firm estimate of pressure at time of formation can be made, and since the pressure-volume-temperature relationships for the three-phase inclusions are unknown, the pressure corrections could not be calculated even if the pressures were known. The possibility that the cavity contents have changed due to movement of material in or out of the cavities cannot be evaluated. Exothermic effects at the faces of the growing crystals are likewise possible.

With so many variables to consider, an unique interpretation of the data in terms of temperatures of crystal formation is impossible. The question is: Can we recognize limits of temperature within which crystallization must have taken place? The discussion that follows is addressed to this question. The "apparently primary" inclusions are the ones chiefly considered; it is assumed for purposes of the discussion that these inclusions either are truly primary or were formed at some stage of the development of the host pegmatite body, hence are significant with respect to conditions of pegmatite formation.

Two-phase inclusions in beryl: Suppose we consider first the apparently primary inclusions in beryl that consist of two phases at room temperature. How can the broad range of temperatures of disappearance of the vapor phase shown by inclusions from the Selden east pegmatite be explained? Values for 86 inclusions are given in Fig. 6. All show disappearance of the vapor phase, and hundreds of others with comparable degrees of filling have been observed in the crystal sections studied. The 36 inclusions of this type in beryl from the Strickland-Cramer pegmatite shown in Fig. 6 show the same behavior. The inclusions therefore cannot fall under Case III of Table 2. It seems unlikely, moreover, that the inclusions fall under Case IV of Table 2, for if this were so, a few, at least, would show lower degrees of filling, and either disappearance of the vapor phase or critical phenomena. For the same reason, differential leakage seems unlikely, for Case IV actually describes the behaviors expectable if inclusions formed under conditions of Case I or Case II have leaked differentially. If, however, the curves for these inclusions are similar to those given for water and a 10 percent KCl solution by Ingerson (1947, Fig. 1), inclusions showing critical phenomena or disappearance of the vapor phase would not be developed unless leakage reached an advanced stage. It is therefore possible either that the inclusions fall under Case I or Case II and that the range of values is due to leakage, or that they fall under Case IV. In either case, the true temperature of formation, uncorrected for pressure, should be nearest the lowest value given by any of the inclusions; namely, 187° C. for beryl from the Selden east pegmatite and 194° C. for beryl from the Strickland-Cramer pegmatite. The same would hold if we regard the variation in temperature as due to exothermic effects at the faces of the growing crystals.

On the other hand, let us consider the possibility that material may be added to the inclusions after their formation, as Kennedy suggested. If the inclusions were formed under conditions of either Case I or Case II, addition of fluid would account for the variations in degree of filling, and also for the fact that none of the inclusions has so low a degree of filling that it shows either critical phenomena or disappearance of the vapor phase. If this is the cause, however, then the true values of temperatures of formation of beryl, uncorrected for pressure, should be nearest the highest temperatures of disappearance; namely, 268° C. for the Strickland-Cramer pegmatite and 385° C. for the Selden pegmatite. From their geologic characteristics, such a marked contrast in temperatures of formation would hardly be expected for the two pegmatites, but the reader may well object that additional temperature studies for the Strickland-Cramer pegmatite might narrow the gap. Kennedy's suggestion is a definite possibility.

Still a third approach is the statistical one advocated by Smith and his co-workers (Smith *et al.*, 1950, p. 587). By this, the temperature of formation of beryl in the Selden east pegmatite, uncorrected for pressure, should be taken as somewhere between 210° and 290° C., the range in which most of the values lie. We might find support for this interpretation in the fact that where groups of inclusions in a crystal give closely similar values, these lie within the range above, and in the further fact that most of the values for beryl from the late-formed units of the Strickland-Cramer pegmatite and the Gillette pegmatite likewise lie in the same range. With so many possible sources of variation in temperatures of disappearance, can we accept a statistical norm? We question this, because the preceding analysis of the possible sources of variation in two-phase liquid inclusions suggests that if a group of inclusions shows a range of degrees of filling, the degree of filling appropriate to the temperature of formation is more likely to be represented by one or the other end of the range than by the median values.

This argument breaks down, however, when we consider that several processes may be jointly responsible for variations in degree of filling of a group of inclusions. One or more processes might increase the degree of filling of some of the inclusions, another process or processes might decrease the degree of filling of still other inclusions. In this case, the proper degree of filling (and temperature of vapor phase disappearance) would be between the two extremes found. But what criteria have we that will enable us to diagnose such cases?

The writers conclude that for the present, the temperature of formation of the beryl in the Selden east pegmatite, uncorrected for pressure, can only be taken as lying within the range 187° C. to 386° C. If the pegmatites formed under rock pressures corresponding to depths of 3 miles, and the compressibilities and thermal characteristics of the inclusions approach those of pure water, the corresponding temperatures of formation would be roughly 272° C. and 600° C. Actually, however, there is no means of determining the true pressure. So far as the data indicate, beryl formed in the Strickland-Cramer and Gillette pegmatites within the same range of temperature as the Selden east pegmatite.

Three-phase inclusions in beryl: If two-phase inclusions present such difficulties, what of the three-phase inclusions, which are even more complex? Let us consider the data for three-phase inclusions in beryl from the Bordonaro north pegmatite, as given in Fig. 11. Various inclusions in beryl from the pegmatite show three types of behavior when heated—(1) filling with a liquid phase, (2) filling with a vapor phase, or (3) disappearance of the phase boundary. Taken individually, these behaviors might result, for (1), under Case I, Case II, or Case IV; for (2) under Case III or Case IV; and for (3) under Case III or Case IV. Our problem, however, is not to account for any single behavior type, but for the entire assemblage. In this connection, the frequency of occurrence of the three behavior types is of interest. The figure shows that of 28 inclusions for which precise data are available, 23 show filling with liquid, three show filling with vapor, and only two show fading of the phase boundary. Inasmuch as all three types are found in a single crystal, and the occurrence of the three types is evidently unrelated to stages of crystal growth, it seems unreasonable to account for the three types on the basis of variations in over-all phase conditions in the mother liquid. Several possibilities remain, however. First, the inclusions may have formed under conditions of Case IV. The true temperature of crystal formation uncorrected for pressure, should then be closest to the lowest value found; namely, 282° C. Second, the variation may be due to differential leakage of inclusions originally formed under Case I or Case II; again the lowest value should be taken. The same would hold true if the variation is due to exothermic effects at the faces of growing crystals; it is necessary to assume, of course, that local heating was enough to produce a vapor phase.

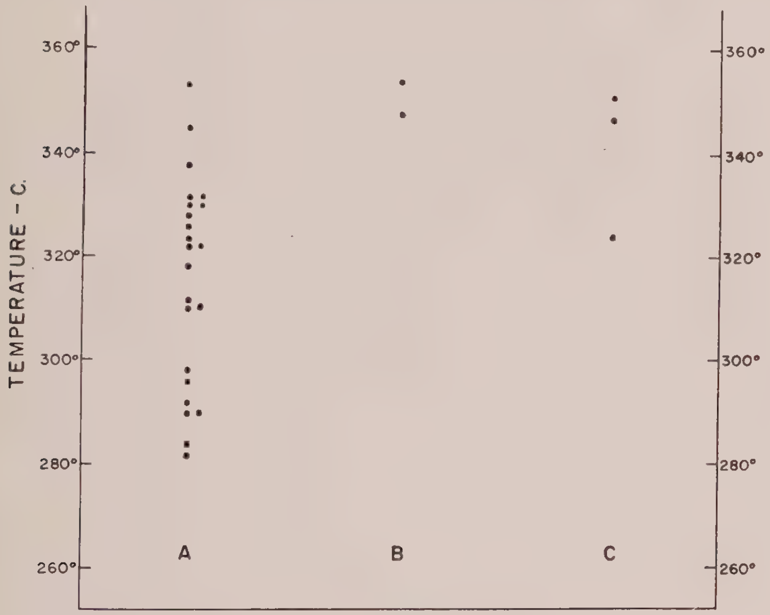

FIG. 11. Temperature data for inclusions in beryl. Bordonaro north pegmatite. Each symbol represents one inclusion. Heating causes disappearance of the vapor phase in inclusions of group *A*, disappearance of the liquid-vapor boundary in inclusions of group *B*, and disappearance of the vapor phase in inclusions of group *C*.

The relationships shown in Fig. 11 are consistent with any of these possibilities. Reference to the work given by Ingerson (1947, Fig. 1), together with rough measurements of degrees of filling in the inclusions in beryl from the Bordonaro pegmatite, indicates the following:

(1) Inclusions showing disappearance of the liquid phase fall within a narrow range of degrees of filling, hence should be relatively few in number, whereas those showing disappearance of the vapor phase fall within a broader range, hence should be relatively numerous. Inclusions showing fading of the phase boundary must have degrees of filling falling within an extremely narrow range, and should be rare.

(2) The temperatures of disappearance of the vapor phase, or of the liquid phase, should not exceed the temperature of disappearance of the phase boundary. This corresponds to Fig. 11, except that one inclusion showed disappearance of the phase boundary at a temperature 6° below the maximum. As has been noted, however, the precise point at which fading of a phase boundary is complete is extremely difficult to determine in such tiny inclusions, and temperatures recorded are considerably less accurate than those recorded for disappearance of liquid or vapor.

The final possibility to be considered is that the variation is due to addition of material to the inclusions. If this be so, the inclusions must have formed under the conditions of Case III [(3)*b*], otherwise we cannot account for the inclusions that show disappearance of the vapor phase or fading of the liquid-vapor boundary. The lowest value of the degree of filling should then be nearest the original value, and the temperature of disappearance of the vapor phase in this inclusion should be nearest the temperature of formation of the inclusion, uncorrected for pressure. For the Bordonaro north pegmatite (Fig. 11) this temperature would be 324° C. Again, as under Case IV, and as in the cases of leakage or exothermic heating, the temperature of disappearance of the vapor-liquid boundary in inclusions showing this behavior should be greater than the temperature of disappearance of the vapor phase in any of the inclusions of group *A*, Fig. 11, or the temperature of disappearance of the liquid phase in inclusions of group *C*.

There is nothing in the characteristics of the inclusions in beryl from the Bordonaro pegmatite that would indicate which of the above is the true explanation. For possible clues, let us consider the data available for three-phase inclusions in beryl in other pegmatites. The data for the Bordonaro south pegmatite are too scant to have much significance, though the fact that results are in the same range as those for the Bordonaro north pegmatite is of interest. For the Case No. 2 pegmatite, (Fig. 9), data for 28 inclusions are available; these inclusions were selected, on the basis of estimates of degree of filling, as representative of hundreds of similar apparently primary inclusions. Temperatures of disappearance range from 292° to 332° C.; this range has its limits within the broader range of values shown by the Bordonaro pegmatite. However, we are here faced with the same difficulties presented by the two-phase inclusions. If Case IV of Table 2 applies, or if leakage is the cause of variation, why is it that none of the inclusions shows disappearance of the liquid phase? Addition of material and exothermic effects remain as possible explanations, the one pointing to acceptance of the highest value as most closely approximating the proper temperature of disappearance, the other pointing to the lowest value. Data for the Gotta-Walden pegmatite present somewhat the same problem as those for the Bordonaro north pegmatite. The upper limit of temperature values is about the same, and all but three values are within the same range. These three are lower, the lowest 238°. Data for the Slocum and Case No. 1 pegmatites are inadequate, though of interest because the values are comparable in general to those obtained for three-phase inclusions in beryl from other pegmatites. Two inclusions in beryl from the Slocum pegmatite, however, give values of 400° C. and 455° C., respectively.

We are thus left to conclude, pending further investigation, that no choice among alternative explanations of the variations in degrees of filling is justified.

Certain comments appear warranted, however, by the available data. The Bordonaro north, Bordonaro south, Case No. 1 and Case No. 2 pegmatites are closely comparable in mineral composition and in structural characteristics, and the beryl studied comes from the same or closely allied mineral assemblages in each pegmatite. The Slocum pegmatite is similar, though the beryl is golden instead of green. The Gotta-Walden pegmatite is structurally different, but the beryl is from the same general mineral assemblage in which beryl occurs in the other five pegmatites. For all six pegmatites, the ranges of values are much the same, and in all six the apparently primary inclusions are of the three-phase type. This suggests (1) that the inclusions are indeed primary, (2) that the temperature data have significance in terms of temperatures of formation, and (3) that the key to the problem is a sound basis for choosing among the various possible causes of variation in the behavior of the inclusions.

Another point worthy of note is the range of critical temperatures indicated for the three-phase inclusions, as follows:

Bordonaro north pegmatite	348°–354° C.
Bordonaro south pegmatite	345° C.
Case #1 pegmatite	311°–316° C.
Slocum pegmatite	329°–347° C.

The ranges are narrow in a given pegmatite, and in all pegmatites the critical temperatures are below those of pure water.

Data for the system H_2O-CO_2 at high temperatures and pressures are not available, hence no basis for calculating pressure corrections for three-phase inclusions exists. However, Kennedy's statement (1950, p. 543)

> "if the solution in the vacuoles contains dominantly dissolved gases, such as CO_2 and H_2O, or volatile fluorides with low critical temperatures, the pressure correction may be greater than for pure water."

is applicable.

Secondary inclusions in beryl: Further work on secondary inclusions in beryl is needed, but a few comments appear warranted at the present stage. The chief points of interest are (1) that results to date give temperature values comparable to those for the apparently primary inclusions, and (2) that if the apparently primary inclusions in a crystal consist of three phases, the secondary inclusions commonly show three phases likewise. The implication is that the secondary inclusions formed from solutions of the same general composition as the apparently primary inclusions and probably only a short time after the host crystals developed; i.e., within the period of pegmatite formation.

Inclusions in quartz: Data for inclusions in quartz from the Case No. 2 pegmatite are impossible to interpret at the present time. The type most likely to be primary, the disseminated negative crystals, gave data so erratic that analysis of the type applied to the inclusions in beryl yields no rational results. Fifteen inclusions showed disappearance of the liquid-vapor phase boundary at temperatures of 293° C. to 357° C. Of these, 14 were found in two specimens of quartz from the core of the pegmatite. One of these specimens contained an inclusion showing disappearance of the vapor phase at 395° C., a temperature higher than the apparent critical temperature of any of the 14 associated inclusions. Variation in the composition of this group of inclusions is therefore suggested.

The contrast in behavior between the disseminated irregular inclusions, in which the vapor phase disappears at temperatures of 124° C. to 292° C., and the disseminated negative crystals is a noteworthy feature, but its meaning is uncertain.

The data for obviously secondary inclusions are too scant to be significant. The great abundance of these inclusions, which are so numerous that they give the specimens a milky, banded appearance, is worth emphasizing. Although they likewise commonly consist of three phases at room temperature, it may be that during their formation the other types of inclusions were irregularly modified.

One of the two specimens of core quartz cited above was Cameron's specimen No. 46-C-3, which was studied earlier by Ingerson (1947, Table 2, No. 5) and Scott (1948, p. 650). Ingerson obtained 155° C. for the probable best value of the temperature of disappearance of the vapor phase in inclusions in this specimen, and Scott obtained a value of 165° C. by decrepitation. Dr. Ingerson kindly returned the specimen for our use, together with his notes. His measurements appear to have been made on disseminated irregular inclusions, most of which are now known to consist of three phases at room temperature. The data here given show that we have been unable to duplicate his results, for although one of our 29 inclusions of this type gave a value of 124° C. and another 160° C., all others gave considerably higher temperatures. In addition, the disseminated negative crystals, which seem more likely to be primary than the others, give values that are much higher than 155° C. Finally, in all specimens of beryl examined by us, including a crystal from the core, the apparently primary inclusions are of the three-phase type and show disappearance of the vapor phase at 292° C. or above, more than 100 degrees above the temperature of 175° C. found by Ingerson for an inclusion in a small crystal of beryl present in specimen 46-C-3.

The value of 165° C. obtained by Scott probably represents bursting of secondary inclusions. These are so exceedingly predominant that their bursting would be expected to mask the bursting of the other types of inclusions. The point of steepening of the decrepitation curve for this specimen (Scott, 1948, Fig. 4*c*) corresponds quite well with the temperature at which we found fracturing and leakage of inclusions to begin when the plates of the quartz were heated on the microscope stage. There is some evidence, however, that the inherent strength of quartz is the controlling factor, rather than the degrees of filling of inclusions.

Composition of the liquid inclusions and the origin of pegmatites: It has generally been assumed that primary liquid inclusions in minerals are accurate samples of the mother liquors from which the host minerals were deposited, and some attention has been given to analysis of inclusions as a guide to the composition of mineral-depositing fluids (e.g., Newhouse, 1932; Buerger, 1932*b*). More recently, Roedder (1951) has discussed the results of analysis of inclusions in pegmatitic quartz. He finds that the inclusions are solutions having concentrations of 10 per cent or more. The ratio of K to Na was found to be about 1 to 1, and the ratio of Cl^-ion to $SO_4^=$ion (sulfide ion would appear here in the analysis) was found to range from 0.5 to 50. The findings suggest moderately concentrated aqueous solutions.

As reported previously (Cameron, Rowe, and Weis, 1951), the fluid inclusions studied, both those in beryl and those in quartz, are highly mobile at room temperatures, and this together with the heating curves for the solutions, the results of freezing experiments, the disappearance of one liquid phase of the three-phase inclusions, and the critical phenomena, suggest strongly that inclusions in beryl in some of the pegmatites are aqueous solutions and that those in beryl and quartz in other pegmatites are similar but richer in carbon dioxide. The inclusions do not have the properties that would be expected of solutions rich in silicates.

The nature of these inclusions has a definite bearing on the problem of the origin of pegmatites. Some of the beryl and quartz studied came from the wall zones and outer intermediate zones of the pegmatites. Let us suppose, for purposes of argument, that the inclusions are truly primary and that they are true samples of the mother fluids from which the pegmatites crystallized. Then if pegmatites crystallize from magma in closed or restricted systems (Cameron, Jahns, McNair, and Page, 1949, pp. 98–105), the primary inclusions in minerals of outer zones should be rich in constituents of the inner zones. Furthermore, as compounds of these constitutents should be relatively insoluble at room temperature, the inclusions should consist of crystals of silicates as well as fluids. If such inclusions are lacking, then the pegmatites must have formed in open systems from fluids of hydrothermal or pneumatolytic character.

This matter merits thorough investigation. It is not a simple one. First, the assumption that the inclusions are representative samples of the mother liquors needs examining. The validity of the assumption for large primary inclusions in minerals is hardly debatable, provided post-depositional modification can be excluded. However, for inclusions measurable in microns, as are all those described in the present paper, this assumption needs checking by laboratory experiments in growing inclusions in polycomponental solutions. Concentration gradients are known to exist in liquids in which crystals are growing (Bunn, 1949). The microns-thick layer of fluid immediately adjacent to the growing face of a crystal may or may not have the same composition as the surrounding liquid.

A second possibility has been pointed out by Professor R. M. Garrels, of Northwestern University (oral communication). Neumann (1948) and Smith (1948*b*) have suggested that immiscibility occurs in pegmatitic liquids, resulting in a predominant silicate-rich liquid and a subordinate water-rich liquid. It is possible that the fluid inclusions studied represent the water-rich liquid.

A third possibility has been suggested by Weis. Our investigation has centered on the fluid inclusions. However, solid inclusions are also present in the crystals of pegmatite minerals. Some are monomineralic and presumably represent accidental inclusions. Others, however, are of unknown composition; they may be mixtures of minerals. If so, have these mixtures crystallized from included mother liquor? At any rate, this matter needs thorough investigation. It is worth remarking that if minute disseminated fluid inclusions are always to be accepted as true samples of mother fluids, then the average granite crystallized from a water-rich solution.

By way of comment on the work of Roedder, it should be pointed out further that bulk analysis of fluid inclusions is valid as a guide to the composition of mineral-forming solutions only if it can be shown that the sole or at least predominant inclusions are of primary origin. In our work on ordinary pegmatite quartz, we have yet to find a specimen in which this is the case.

CONCLUSIONS

The following conclusions and comments appear indicated by the studies discussed above:

(1) Inclusions in quartz and beryl in the pegmatites studied are of three principal types—(1) disseminated negative crystals or tubular inclusions, which in beryl are crystallographically oriented; (2) disseminated rounded or irregular inclusions; (3) inclusions along healed fractures. Inclusions of type (1) are probably primary, those of type (2) may be either primary or secondary, those of type (3) are secondary.

(2) Inclusions of type (1) in the Selden east pegmatite and the Strickland-Cramer pegmatite consist of two phases at room temperature. These inclusions appear to be aqueous solutions with concentrations ranging from 2.7 to 6.5 molar. The vapor phases in various inclusions from the Selden east pegmatite disappear at temperatures ranging from 187° C. to 386° C. Pressure corrections equivalent to a rock load of 15,000 feet would raise these values to 272° C. to 600° C., a range for the possible temperature of crystallization that is unsatisfactorily broad. The range of values, uncorrected, is 196° to 268° C. for inclusions in beryl from the Strickland-Cramer pegmatite.

(3) Inclusions of type (1) in beryl from six pegmatites consist of two liquid phases and a vapor phase at room temperature. One liquid phase disappears at about 31° C., hence appears to consist essentially of liquid carbon dioxide. With further heating, the vapor phase in some of the inclusions disappears, at 238° C. to 356° C. in various inclusions. In others the liquid phase dis-

appears, at 310° C. to 455° C. In still others the liquid-vapor phase fades and disappears, at 329° C. to 348° C. Inclusions of all three types are present in beryl from three of the pegmatites.

(4) Limited data suggest that inclusions of types (2) and (3) are comparable to those of type (1) in composition and behavior and hence were probably also formed during the period of pegmatite crystallization.

(5) Inclusions in quartz have proved unsatisfactory for temperature studies, because most of them fracture and leak when heated. Inclusions of type (1) are uncommon. Most show critical phenomena when heated and are of no value as indicators of temperatures of crystal formation. Most, perhaps all, such inclusions in quartz from the Case No. 2 pegmatite consist of three phases at room temperature. Inclusions of type (2) are more common. Those in quartz from the Case No. 2 pegmatite consist of three phases at room temperature, show disappearance of one liquid phase at about 31° C., and filling with liquid at 124° C. to 292° C. Inclusions of type (3) greatly predominate over the other types, but few data have been obtained, owing to fracture and leakage of the inclusions when heated.

(6) Decrepitation curves for quartz in pegmatites are significant only when bursting can be correlated precisely with the various types of inclusions, when the inclusions can be shown to consist of aqueous solutions, when premature bursting or leakage can be shown not to take place, and when the effects of primary inclusions are not masked by those of secondary inclusions. Our study suggests that these conditions will prove difficult to satisfy.

(7) Analysis of potential causes of variations in temperatures of disappearance of vapor and liquid phases and liquid-vapor boundaries indicates that a unique interpretation of the temperature data obtained is impossible at present. The actual temperatures of formation, uncorrected for pressure, may lie at one end or the other of the ranges of temperature values obtained, and there is no assurance that the statistical norm is significant.

(8) Absolute proof that inclusions in pegmatite minerals are primary is difficult to obtain, particularly for inclusions in quartz.

(9) True temperatures of pegmatite crystallization are probably considerably higher than temperatures of phase or phase boundary disappearance in liquid inclusions, but until some unequivocal means of determining pressures at time of crystallization can be devised, estimated corrections for pressures are little more than guesses based on geological possibilities. Owing to this, to

uncertainties as to the compositions of liquid inclusions, and to lack of *P-V-T* data for systems other than water at high temperatures and pressures, the liquid inclusion method is not at present suited to accurate determination of temperatures of mineral formation in the pegmatites studied.

(10) Several questions must be answered before inclusions in pegmatite minerals can be accepted as true samples of the liquids from which pegmatites crystallize.

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FAHEYITE, A NEW PHOSPHATE MINERAL FROM THE SAPUCAIA PEGMATITE MINE, MINAS GERAIS, BRAZIL*

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Abstract

Faheyite, a new mineral from the Sapucaia pegmatite mine, Minas Gerais, Brazil, has the composition $(Mn,Mg,Na)Be_2Fe_2(PO_4)_4 \cdot 6H_2O$. It occurs in vugs as white, bluish-white, or brownish-white tufted fibers that coat other minerals, such as muscovite, quartz, variscite, and frondelite.

The powder pattern of faheyite has been completely indexed on the basis of a primitive hexagonal cell with dimensions: $a_0 = 9.43$ Å and $c_0 = 16.00$ Å. The powder film was compared with a rotation picture of a fiber taken in the powder camera, as an aid to indexing the powder film. *X*-ray powder photographs show strong lines at 5.72, 7.28, 3.243, 3.087, 3.031, and 3.958 Å. The measured specific gravity is 2.660, and the specific gravity calculated from *x*-ray data is 2.670. Indices of refraction are $\omega = 1.631$ and $\epsilon = 1.652$.

Chemical analysis gave the following percentages: insoluble 9.44, P_2O_5 38.11, Fe_2O_3 21.42, Al_2O_3 0.10, Mn_2O_3 none, FeO none, BeO 7.26, MnO 5.99, MgO 1.14, Na_2O 0.84, K_2O trace, F trace, H_2O 14.90; total 99.20.

The mineral is named in honor of Joseph J. Fahey, geochemist of the U. S. Geological Survey.

Introduction

In the period 1943 to 1945, William T. Pecora, of the U. S. Geological Survey, and A. L. de M. Barbosa, of the Departamento Nacional da Producção Mineral, Brazil, made several examinations of the Sapucaia pegmatite mine, near Conselheiro Pena, Minas Gerais, Brazil, and collected a representative suite of minerals for later study. These minerals have been under investigation in the laboratory of the U. S. Geological Survey since 1947. The present paper describes a new phosphate mineral, the second new mineral from this locality (Lindberg, 1949). A complete description of the pegmatite and its minerals will appear in a forthcoming paper.

Faheyite is a hydrous beryllium-manganese-iron phosphate with the formula $(Mn,Mg,Na)Be_2Fe_2(PO_4)_4 \cdot 6H_2O$. It is named in honor of Joseph J. Fahey, of the Geochemistry and Petrology Branch, U. S. Geological Survey, in recognition of his contributions to the chemistry of minerals and of the guidance he has given to younger chemists engaged in the analytical chemistry of minerals.

Occurrence and Physical Properties

Faheyite occurs in vugs as white, bluish-white, or brownish-white fibers coating other minerals. Botryoidal masses of fibers completely enclose euhedral quartz crystals or are attached to surfaces of muscovite sheets. Flat rosettes of fibers lie between sheets of muscovite, and tufts of fibers occur on crystals of variscite, on botryoidal frondelite, and between layers of frondelite.

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* Publication authorized by the Director, U. S. Geological Survey.

Individual fibers of faheyite average about 0.08 mm. in length and 0.01 mm. in thickness. The fibers usually grow normal to the surfaces of other minerals and may be singly terminated by pyramid faces.

Faheyite is uniaxial (+); $\omega = 1.631$ and $\epsilon = 1.652$; $\epsilon - \omega = 0.021$; elongation is parallel to the *c*-axis; cleavage, perfect and parallel to *c*-axis.

The specific gravity was measured on the sample used for analysis by means of an Adams Johnston pycnometer of fused silica; it was found to be 2.660 at 4° C. This sample contains 9.44% insoluble matter, which consists chiefly of quartz, with a minor amount of muscovite. The specific gravity was not corrected for the insoluble matter, since the specific gravity of faheyite is essentially the same as that of quartz. The specific gravity calculated from *x*-ray data is 2.670.

CHEMICAL COMPOSITION

Faheyite is essentially a hydrous beryllium-manganese-iron phosphate, with the formula $(Mn,Mg,Na)Be_2Fe_2'''(PO_4)_4 \cdot 6H_2O$. Its chemical analysis, ratios, calculated equivalents, and atoms per unit cell are listed in Table 1. The number of atoms of each type are calculated by multiplying the equivalents by 0.01 to convert from a percentage to a fractional scale, and by 1975, the unit-cell weight expressed on a chemical scale against O = 16. The role that sodium plays in the mineral is uncertain; but sodium probably substitutes for divalent manganese, which has a similar ionic radius. If such a substitution occurs, the number of atoms in the unit cell in the manganese position (Mn, Mg, Na) becomes a whole number.

At the Sapucaia pegmatite mine, beryllium occurs in small quantities in phosphate minerals associated with faheyite. Spectrographic work shows 0.0X% Be in frondelite and 0.0X% to 0.X% Be in variscite.

Faheyite is slowly dissolved by hot dilute HCl, HNO_3, and H_2SO_4.

X-RAY DATA

This beryllium-manganese-iron phosphate occurs as minute fibers; these failed to give any reflection in single crystal studies, and the Laue symmetry is not known. Carefully selected parallel bundles of fibers gave good rotation photographs around the *c*-axis, although the spots representing various families of planes were spread through a small arc. Every Weissenberg photograph taken showed a completely random orientation perpendicular to the fiber axis.

TABLE 1. CHEMICAL ANALYSIS AND FORMULA OF FAHEYITE

	Analysis	Recalculated after Deducting Insoluble	Ratios	Oxygen Equivalent	Metal Equivalent	Atoms: Metals and Water per Unit Cell
Insol.*	9.44					
P_2O_5	38.11	42.08	0.2964	1.4821	0.5928	11.71
Fe_2O_3	21.42	23.65	0.1481	0.4443	0.2962	5.85
Al_2O_3	0.10	0.11	0.0011	0.0032	0.0021	0.04
Mn_2O_3	None	None				
FeO	None	None				
BeO	7.26	8.02	0.3205	0.3205	0.3205	6.33
MnO	5.99	6.61	0.0932	0.0932	0.0932	1.84
MgO	1.14	1.26	0.0313	0.0313	0.0313	0.62
Na_2O	0.84	0.93	0.0150	0.0150	0.0300	0.59
K_2O	Trace	Trace				
F	Trace	Trace				
H_2O	14.90	16.45	0.9131			18.03
	99.20	99.11				

M. L. Lindberg, analyst. Analysis on 1.2 grams.

Alkali determination by flame photometer by E. Nygaard. A spectrogram by K. J. Murata gives in addition 0.X% Ca, 0.0X% Ti, Zn, and Pb, and 0.00X% Co, Sr, and Ba. Elements looked for but not found: Cu, Ag, Bi, As, Sb, Sn, Cd, Tl, Ce, In, Ga, Mo, W, Bi, V, Cr, Zr, Cb, La, Y, Sc, and B.

* Insoluble consists of quartz and muscovite.

The rotation photographs were used in conjunction with the powder photographs as an aid to indexing the latter. In a hexagonal mineral

$$\frac{1}{d^2_{hkl}} - \frac{l^2}{c^2} = \frac{4}{3}\ \frac{(h^2 - hk + k^2)}{a^2};$$

in faheyite $1/d^2_{hkl}$ was obtained by measuring the powder photograph, and l^2/c^2 for each *hkl* plane was obtained in the following manner: Rotation photographs around the fiber *c*-axis were taken both in the powder and in the rotation cameras. The rotation photographs from the powder camera were aligned with the powder pictures from the same camera, and the *l*-index corresponding to a given line on the powder photograph was found. Photographs taken with iron radiation gave indices of $l=0$, 1, and 2 (Fig. 1); photographs taken with copper radiation gave indices of $l=0$, 1, 2, and 3. The higher values of l were obtained by the use of the Bernal chart in conjunction with the regular rotation photograph. For points outside the sphere of reflection, that rotate

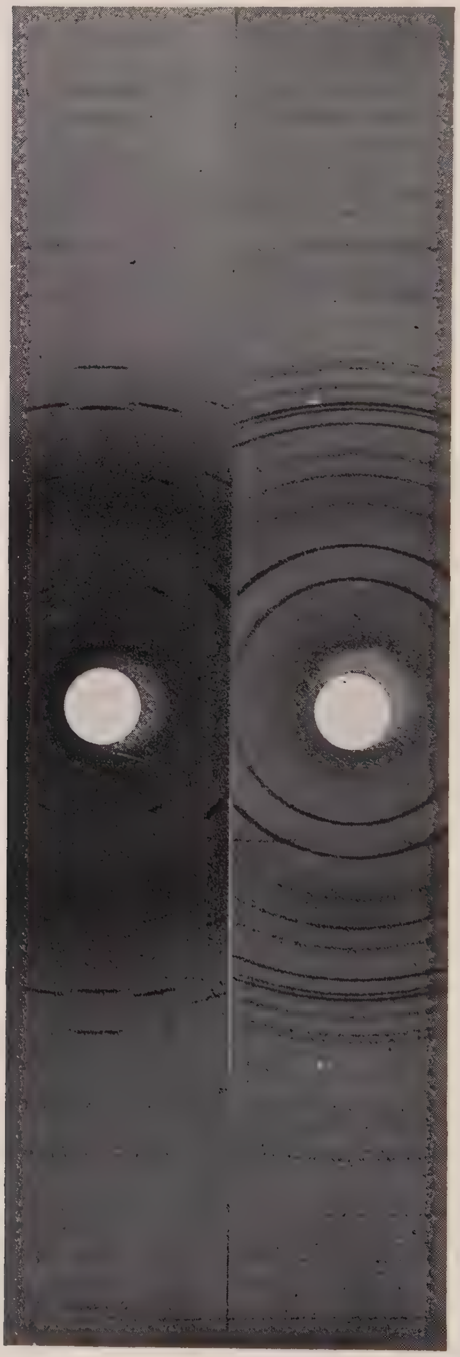

FIG. 1. Powder and rotation photographs of faheyite, powder camera. (Iron radiation, manganese filter.)

about the c-axis, the l-index was found by calculation after the value for a_0 had been established.

To find a_0 the equation

$$\frac{1}{d^2_{h0l}} - \frac{l^2}{c^2} = \frac{4}{3}\,\frac{h^2}{a^2}$$

was solved for 21 reflections indexed as 101, 102, 103, 104, 105, 107, 201, 203, etc., and an average value of 0.014984 for $4/3a^2$ was obtained. This value was used to find the calculated values of d_{hkl} according to the equation:

$$\frac{1}{d^2_{hkl}} = 0.014984\,(h^2 + hk + k^2) + \frac{l^2}{c^2}\cdot$$

This solves to $d_{100} = 8.17$ Å and $a_0 = 9.43$ Å. The value for c_0 obtained from the layer line separation on the rotation pattern was in good agreement with the c_0 derived from the 002 and 004 spacings on the powder photograph, and there were no consistent variations between measured and calculated $\frac{1}{d^2_{hkl}}$ for higher values of l^2/c^2. The cell volume is 1233 Å^3.

All spots were indexed, and no systematic absences were observed; the pattern indicates a primitive unit cell. Calculated and measured values for d are given in Table 2.

CONCLUSIONS

Faheyite is a hydrous phosphate that formed late in the mineral sequence of the Sapucaia granitic pegmatite. Other phosphate minerals in this pegmatite are triphylite, heterosite, vivianite, frondelite, roscherite, childrenite, hureaulite, apatite, and variscite. Faheyite has no apparent structural relationship to these or to other phosphate minerals recorded in the available literature.

ACKNOWLEDGMENTS

The writers are indebted to colleagues in the Geochemistry and Petrology Branch, U. S. Geological Survey, particularly to William T. Pecora for providing the faheyite material and to George T. Faust and Charles L. Christ for their valuable suggestions. The Department of Geology, Dartmouth College, kindly donated additional material to supplement the original material from which faheyite was recovered for analysis.

REFERENCE

LINDBERG, Marie Louise (1949), Frondelite and the frondelite-rockbridgeite series: *Am. Mineral.*, **34**, 541–549.

Manuscript received Aug. 24, 1952.

Table 2. *X*-ray Powder Spacing Data for Faheyite

(Iron radiation, manganese filter, $\lambda=1.9373$)
$a_0=9.43$ Å, $c_0=16.00$ Å, $V=1233$ Å^3

Intensity	*d* (meas.) Å	Indices		*d* (calc.)	Indices		*d* (calc.)	Indices		*d* (calc.)
		hkl	*hkīl*		*hkl*	*hkīl*		*hkl*	*hkīl*	
½	8.00	002*	0002*	8.00						
9	7.28	101	$10\bar{1}1$	7.28						
10	5.72	102	$10\bar{1}2$	5.72						
½	4.68	110	$11\bar{2}0$	4.72						
2	4.53	111	$11\bar{2}1$	4.52						
½	4.47	103	1013	4.47						
½	4.02	004*	0004*	4.00						
5	3.962	201	2021	3.958						
2	3.636	202	$20\bar{2}2$	3.638						
2	3.591	104	$10\bar{1}4$	3.593						
½	3.536	113	$11\bar{2}3$	3.533						
6	3.244	203	$20\bar{2}3$	3.243						
½	3.173	005*	0005*	3.200						
6	3.085	210	$21\bar{3}0$	3.087						
6	3.029	211	$21\bar{3}1$	3.031						
2	2.986	105	$10\bar{1}5$	2.981						
2	2.877	212	$21\bar{3}2$	2.880						
2	2.856	204	$20\bar{2}4$	2.858						
3	2.724	300	$30\bar{3}0$	2.723						
3	2.673	213	$21\bar{3}3$	2.672	006*	0006*	2.666			
½	2 583	302	$30\bar{3}2$	2.578						
½	2.531	106*	$10\bar{1}6$*	2.535						
½	2.449	214	$21\bar{3}4$	2.444						
½	2.424	303	$30\bar{3}3$	2.425						
½	2.365	220	$22\bar{4}0$	2.364						
½	2.325	116	$11\bar{2}6$	2.322						
1	2.264	310	$31\bar{4}0$	2.266	222	$22\bar{4}2$	2.262			
1	2.246	311	$31\bar{4}1$	2.243						
½	2.202	107*	$10\bar{1}7$*	2.200						

TABLE 2—(*Continued*)

Intensity	d (meas.) Å	Indices		d (calc.)	Indices		d (calc.)	Indices		d (calc.)
		hkl	$hk\bar{i}l$		hkl	$hk\bar{i}l$		hkl	$hk\bar{i}l$	
½	2.182	312	$31\bar{4}2$	2.180						
1	2.160	223	$22\bar{4}3$	2.157						
½	2.080	313	$31\bar{4}3$	2.086						
1	2.060	117*	$11\bar{2}7$*	2.057						
2	2.027	224	$22\bar{4}4$	2.031	401	$40\bar{4}1$	2.026	216	$21\bar{3}6$	2.018
1	1.973	314	$31\bar{4}4$	1.971						
1	1.944	108*	$10\bar{1}8$*	1.943						
2	1.906	403	$40\bar{4}3$	1.907	306	$30\bar{3}6$	1.905			
½	1.859	321	$32\bar{5}1$	1.861						
½	1.849	315	$31\bar{4}5$	1.849						
2	1.824	322	$32\bar{5}2$	1.825						
½	1.786	410	$41\bar{5}0$	1.783						
1	1.769	411	$41\bar{5}1$	1.771	323	$32\bar{5}3$	1.768			
1	1.740	412	$41\bar{5}2$	1.740	109*	$10\bar{1}9$*	1.737			
1	1.721	405	$40\bar{4}5$	1.722						
2B	1.693	324	$32\bar{5}4$	1.697	413	$41\bar{5}3$	1.691			
2B	1.629	500	$50\bar{5}0$	1.634	501	$50\bar{5}1$	1.625	406	$40\bar{4}6$	1.622[a]
2B	1.573	330	$33\bar{6}0$	1.576						
½	1.559	503	$50\bar{5}3$	1.562	415	$41\bar{5}5$	1.557			
½	1.538	332	$33\bar{6}2$	1.543						
½	1.529	326	$32\bar{5}6$	1.533						
½	1.513	333	$33\bar{6}3$	1.508	11.10*	$11\bar{2}.10$*	1.515			
1B	1.487	20.10*	$20\bar{2}.10$*	1.490	416	$41\bar{5}6$	1.483			
1B	1.464	510	$51\bar{6}0$	1.467	334	$33\bar{6}4$	1.463	511	$51\bar{6}1$	1.462
½	1.441	244	$24\bar{6}4$	1.440						
½	1.407	513	$51\bar{6}3$	1.414	415	$41\bar{5}5$	1.407			

* This index corresponds to a point that was outside the sphere of reflection when the crystal was rotated about *c*-axis and hence was not observed on the rotation picture.

The powder film was calibrated against shrinkage by addition of Ag metal to the spindle of faheyite. This calibration was made at the suggestion of C. L. Christ. The agreement of calculated and observed d values show that any errors involved in determining the cell edges are less than the errors reading 2θ to ±0.1 mm. (camera diameter 114.59 mm.).

TABLE 2—(*Continued*)

Intensity	*d* (meas.) A	Indices		*d* (calc)	Indices		*d* (calc.)	Indices		*d* (calc.)
		hkl	*hkīl*		*hkl*	*hkīl*		*hkl*	*hkīl*	
½	1.388	11.11*	$11\bar{2}.11$*	1.390	506	$50\bar{5}6$	1.393			
½	1.376	514	$51\bar{6}4$	1.377						
½	1.366	20.11*	$20\bar{2}.11$	1.370						
½	1.357	601	$60\bar{6}1$	1.357						
1B	1.333	341	$34\bar{7}1$	1.338	246	$24\bar{6}6$	1.366	515	$51\bar{6}5$	1.334[b]
½	1.314	603	$60\bar{6}3$	1.319	10.12*	$10\bar{1}.12$*	1.316			
½	1.300	343	$34\bar{7}3$	1.302						
½	1.285	516	$51\bar{6}6$	1.285						
3	1.268	253	$25\bar{7}3$	1.270	20.12*	$20\bar{2}.12$	1.267			
½	1.251	605	$60\bar{6}5$	1.253						
½	1.241	611	$61\bar{7}1$	1.242	345	$34\bar{7}5$	1.238			
½	1.230	612	$61\bar{7}2$	1.231						
½	1.221	248*	$24\bar{6}8$	1.222						
½	1.210	32.10	$32\bar{5}.10$	1.217						
½	1.173	441	$44\bar{8}1$	1.176						
½	1.139	703	$70\bar{7}3$	1.140	353	$35\bar{8}3$	1.140			
½	1.129	616	$61\bar{7}6$	1.129						
½	1.119	262	$26\bar{8}2$	1.122	704	$70\bar{7}4$	1.120	354	$35\bar{8}4$	1.120[c]
½	1.104	20.14*	$20\bar{2}.14$*	1.101						
½	1.085	50.11*	$50\bar{5}.11$*	1.086						
½	1.080	710	$71\bar{8}0$	1.082	711	$71\bar{8}1$	1.080	51.10*	$51\bar{6}.10$*	1.081
½	1.068	41.12	$41\bar{5}.12$	1.068	33.11	$33\bar{6}.11$	1.068			
½	1.042	451	$45\bar{9}1$	1.044	714	$71\bar{8}4$	1.044	11.15*	$11\bar{2}.15$*	1.040
½	1.025	453	$45\bar{9}3$	1.026						
½	1.009	363	$36\bar{9}3$	1.010	24.12*	$24\bar{6}.12$	1.009			

Additional possible reflections:

a	209*	$20\bar{2}9$*	1.629
b	00.12*	000.12*	1.333
c	33.10*	$33\bar{6}.10$*	1.121

B broad reflection.

QUANTITATIVE ANALYSIS OF ENDELLITE, HALLOYSITE AND KAOLINITE BY DIFFERENTIAL THERMAL ANALYSIS

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Abstract

A determination of the relative amounts of endellite versus kaolinite and/or halloysite in a sample is obtained by the differential thermal analysis technique. When a sample containing a mixture of these clay minerals is saturated with ethylene glycol and a differential thermal curve obtained, the kaolinite and/or halloysite gives the usual endothermal reaction at approximately 575° C., whereas the endellite-ethylene glycol complex results in a new endothermal peak at 500° C. Standard reference curves for a series of mixtures of the endellite complex and kaolinite (or halloysite) were obtained and a graph made utilizing the ratios of the two peak heights to give a quantitative measure of the minerals. Values obtained were within 3% of the weighed amount of the given clay in the standard mixtures. Three dissimilar differential thermal analysis units were used to test the method and the effect of impurities on the curves was evaluated.

Introduction

In the investigation of clays containing the members of the kaolin group—endellite, halloysite, and kaolinite, it has been possible only to estimate the amounts of the three similar varieties by the usual techniques of clay mineral analysis. A partial solution to the problem of obtaining a quantitative mineralogical analysis is offered by a differential thermal technique which gives the amount of endellite versus halloysite and/or kaolinite in a sample.

As Bradley (1946) and MacEwan (1948) have shown, various polar organic liquids will replace the interlayer water in endellite. When differential thermal analyses were run on certain of these endellite-organic liquid complexes, such as endellite with ethylene glycol, di- and triethylene glycol, and glycerine, the endothermal peak usually occurring at approximately 575° C. was lowered to 500° C. Since this did not happen with kaolinite or halloysite, which do not form the complex, it became evident that this phenomenon might be used as the basis of a method for determining the amount of endellite versus kaolinite and/or halloysite in a sample.

Obtaining Standard Reference Patterns

Endellite from the Dragon Mine, Eureka, Utah, and ethylene glycol were used to form the complex by shaking the clay (aggregate size between 80 and 200 mesh) in an excess of the glycol for five minutes. The suspension was centrifuged at 3,000 rpm and the liquid decanted. Excess ethylene glycol was removed by working the complex on retentive filter paper. Halloysite produced by dehydrating the endellite at 110° C. was treated in exactly the same manner so that proportional weights of halloysite and endellite-glycol complex were obtained. (Placing the halloysite and endellite complex in a controlled atmosphere might give better standard mixtures, but the stability of the complexes has not yet been worked out and there is the possibility of some of the endellite changing to halloysite. This remains, however, as a possible future refinement of the method.) Ethylene glycol was added to the weighed mixture which was then worked into the consistency of a thick paste and run on the differential thermal unit.

The apparatus used for obtaining the patterns shown in Fig. 1 is similar to that described by Kerr and Kulp (1948). The sample holder is a nickel block with a $\frac{1}{4}''$ dia. × $\frac{3}{8}''$ hole to contain the sample and covered with a nickel lid. Calcined alumina is used as reference material. The sample block is placed on top of a vertically mounted muffle tube over which a furnace is lowered. The rate of heating is 10° C. min. and the differential thermal effect is recorded on an automatically recording Leeds and Northrup Speedomax potentiometer using a sensitivity scale on the preamplifier providing a full-scale deflection of 0.6 millivolts for chromel-alumel thermocouples on the Speedomax chart. The chart moves at a rate of 6 inches/hr. to give a temperature interval of 100° C./inch.

A series of twenty-one mixtures at 5% intervals was prepared and run, and the curves in Fig. 1 are the endothermal doublets obtained for the standard mixtures. The endothermal peak at 500° C. represents the structural water leaving the endellite complex while the 575° C. peak represents structural water leaving the halloysite. There is a drift to the exothermal side due to the smaller specific heat capacity of the sample than of the standard, so the base line is drawn as in Fig. 3.

Kaolinite from Langley, South Carolina, was substituted for halloysite and mixtures of kaolinite and the endellite complex were also run on the *DTA* unit. Since it was found that kaolinite gave the same peak intensity as halloysite, this method gives the relative amounts of endellite versus halloysite and/or kaolinite in a sample. Where all three minerals are present in a sample other techniques must be used to obtain the proportion of kaolinite and halloysite. Recent base-exchange capacity determinations have suggested a method of quantitatively determining the amounts of kaolinite versus halloysite and/or endellite in mixtures. In special cases where the kaolinite occurs as coarser vermicular crystals, either point counts can be made of the thin sections (Chayes, 1949), or the kaolinite crystals may be fractioned off leaving the fine fractions consisting only of halloysite and endellite.

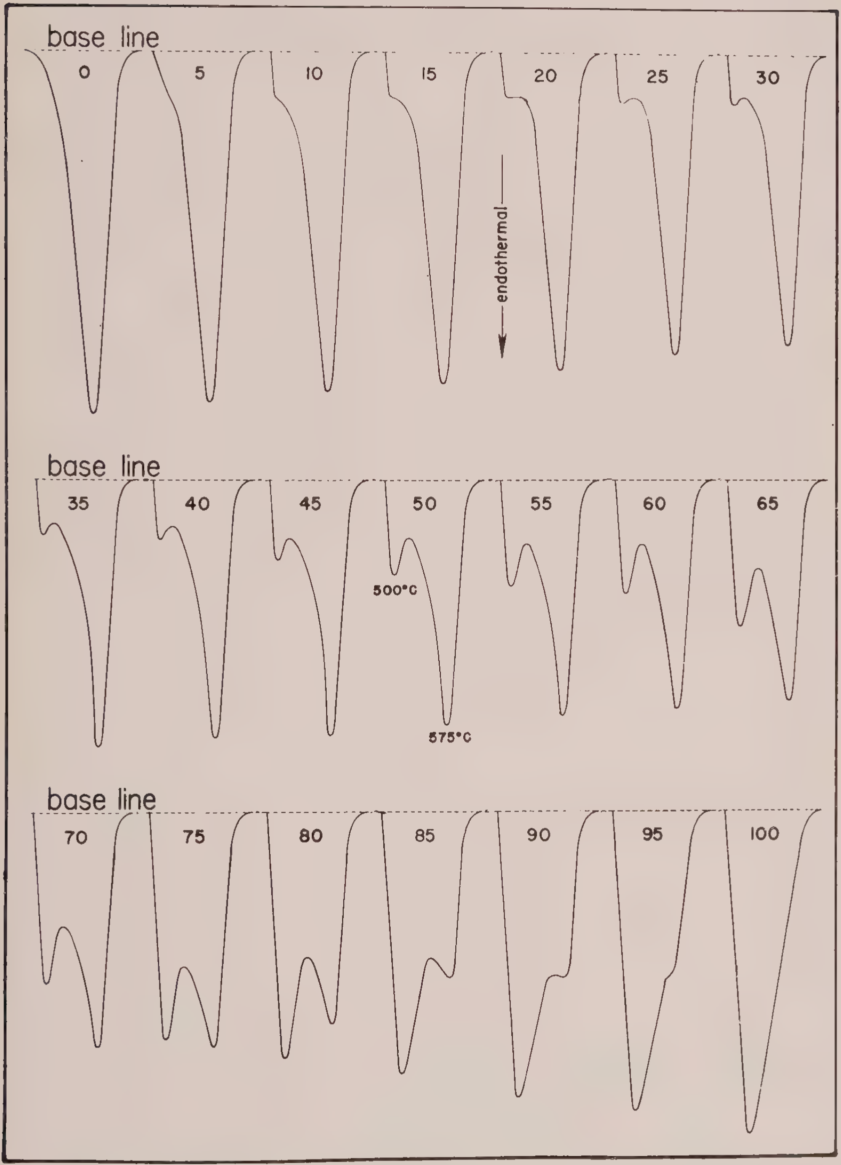

Fig. 1. Differential thermal patterns of standard mixtures containing endellite and halloysite. Percentages are given for content of endellite.

The patterns in Fig. 1 are used for direct comparison with those obtained from unknown mixtures and are especially useful in the low and high percentage mixtures.

Graphical Solution

For the mixtures between 20% and 80% of one component, it is preferable to use the graph shown in Fig. 2. Here the ratio of the height of the 500° C. peak from base line (*E*) to the height of the 575° C. peak (*H*) is plotted along the logarithmic ordinate and the percentage of endellite

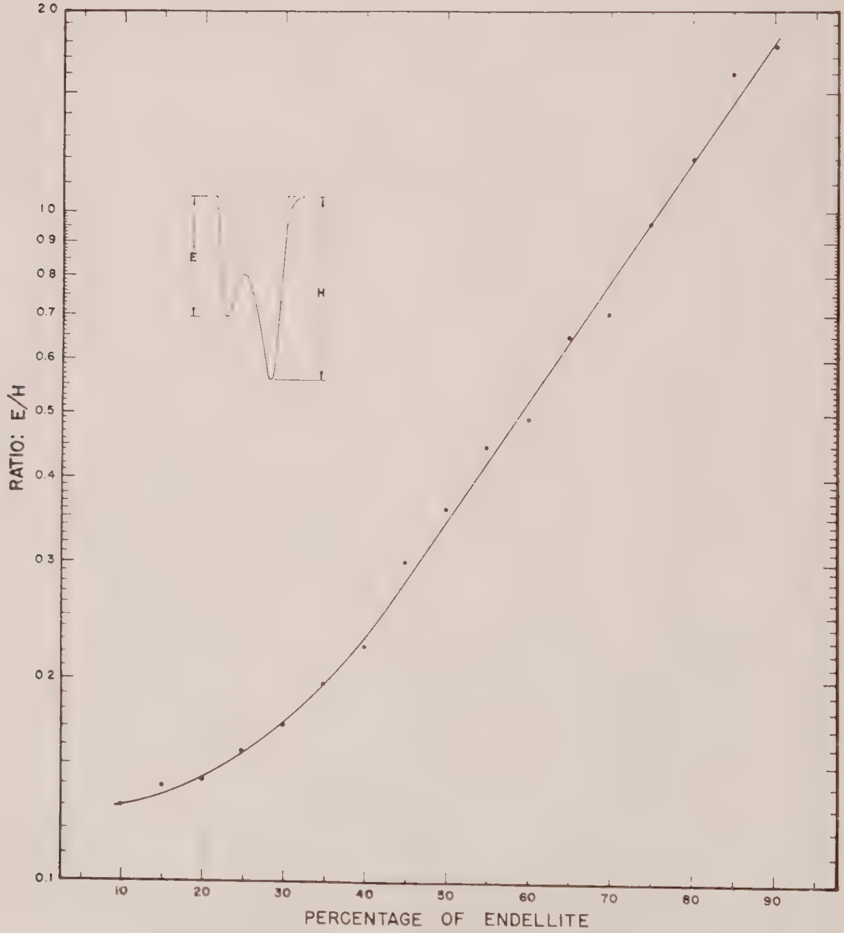

Fig. 2. Graphical solution for obtaining the percentage of endellite in a mixture with halloysite and/or kaolinite.

in the mixture along the abscissa. The ratios as obtained from the standard patterns are as follows:

% Endellite	*Ratio:E/H*	*% Endellite*	*Ratio:E/H*
10	.130	55	.450
15	.138	60	.496
20	.142	65	.660
25	.157	70	.720
30	.172	75	.980
35	.198	80	1.16
40	.226	85	1.60
45	.302	90	1.78

For manually recording potentiometers, where a continuous pattern is not obtained, the graphical solution, which necessitates only the measurement of maximum peak heights, is superior to direct comparison with the reference curves. The over-all accuracy of both methods using our apparatus is ±3% and might be improved by refinement of technique in preparing the standards. It is essential that the base line be established as in Fig. 3. An exothermal peak occurs just before the 500° peak but the return of the 575° peak is to base line.

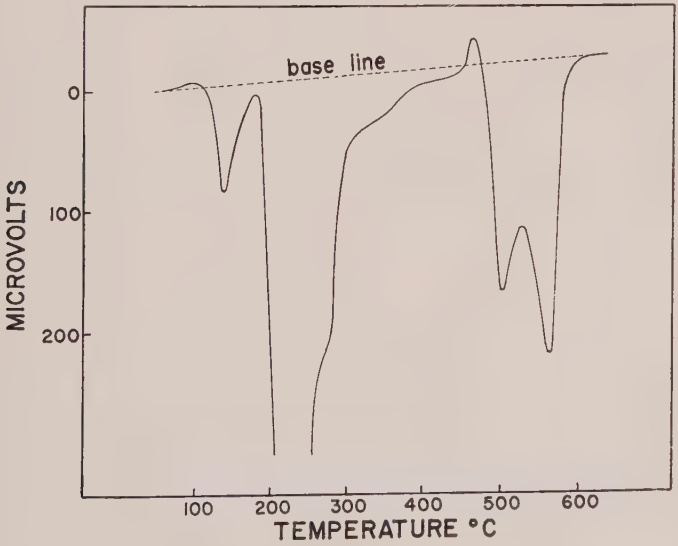

FIG. 3. Differential thermal analysis pattern of a mixture containing 72% endellite-ethylene glycol complex and 28% halloysite.

PREPARATION OF UNKNOWN FOR ANALYSIS

The unknown sample should be mulled in water so the clay aggregate will wet sieve through 80 mesh and be caught on 200 mesh. The clay-water suspension is centrifuged in 15 ml. tubes and the clear water decanted. A large excess of ethylene glycol is added to the wet clay and the tube vigorously shaken for five minutes. This is centrifuged at 3,000 rpm for about fifteen minutes and the clear ethylene glycol decanted. The resulting clay-glycol suspension is of the right consistency for differential thermal analysis. (If no centrifuge is available the excess glycol may be removed by working the mixture on filter paper.) If the clay aggregate size is below ten microns, the suspension after centrifugation is not sufficiently compact and the ethylene glycol must be evaporated in an oven until the suspension has the proper consistency.

Effect of Impurities

Since samples to be analyzed for content of the kaolin minerals usually contain other minerals, the effect of impurities on the curves was studied. A mixed layer dioctahedral 2:1 clay, an expanded dioctahedral 2:1 clay, quartz and mica were selected as minerals commonly associated with kaolinite and halloysite, and mixtures of these with the endellite complex were run on the differential thermal analysis unit.

The mixed layer clay was a K-bentonite from Oak Hall, Pennsylvania, containing 20% expanded and 80% non-expanded layers (Weaver and Bates, 1952). A mixture of 25% of this material and 75% endellite complex yielded a curve (Fig. 4*a*) which is similar to that obtained for 95% endellite and 5% halloysite. A mixture of 50% mixed layer clay and 50% endellite complex gave curve 4*b* which is similar to 90% endellite and 10% halloysite. A mixture of 50% mixed layer clay, 25% endellite complex, and 25% halloysite gave a curve similar to 42% endellite and 58% halloysite. 10% mixed layer clay in the mixture had no evident effect on the shape of the curve.

The expanded 2:1 clay (montmorillonite from Polkville, Mississippi) had no effect on the curves except to decrease the peak heights.

Vein quartz from Washington, D. C. gave a sharp peak at 573.5° C. which was used as the temperature reference. When ground to 200 mesh and mixed with the endellite complex the quartz had the same effect as the mixed layer clay in causing the halloysite peak to be present to a degree equal to about one-fifth the amount of impurity present in the sample. Curves of a series of quartz-endellite complex mixtures are shown in Fig. 4*c*, *d*, *e*.

Muscovite ground to 200 mesh had the same effect on the curves as the mixed layer clay and quartz.

The effect of impurities, as shown by these curves, is to dampen the peak intensities and produce an erroneous halloysite or kaolinite peak which in magnitude is approximately equal to one-fifth of the amount of impurity present. Less than 10% of impurity in the sample will only

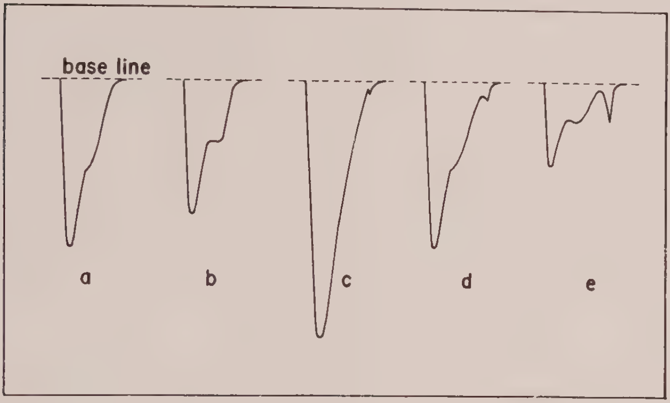

Fig. 4

a. 25% illite and 75% endellite-ethylene glycol complex.
b. 50% illite and 50% endellite-ethylene glycol complex.
c. 10% quartz and 90% endellite-ethylene glycol complex.
d. 25% quartz and 75% endellite-ethylene glycol complex.
e. 50% quartz and 50% endellite-ethylene glycol complex.

cause dampening. Because of its ability to retain the glycol, the expanded 2:1 clay is the only commonly associated mineral which in amounts greater than 10% has no effect on the shape of the curve. The erroneous halloysite or kaolinite peak is produced in the case of the other impurities presumably because there is insufficient glycol retained in the mixture and a small amount of the endellite changes to halloysite. This will also happen if a cover is not placed over the sample, or if the amount of sample is too small or not of the correct consistency.

Analysis on Other Differential Thermal Analysis Units

To test the application of the method using dissimilar differential thermal analysis units, samples were run on Mr. R. M. Gruver's apparatus in the Division of Ceramics and on Dr. C. D. Jeffries' unit in the Department of Agronomy. These units are described in the respective papers by these authors. Mr. Gruver's unit consists of a horizontally mounted furnace and sample holders which are 1.3 ml. platinum crucibles with covers. The rate of heating is automatically controlled and the differential thermal effects automatically recorded on a Brown potentiometer. Dr. Jeffries uses a large nickel block with cover which holds a two gram sample and is placed in a horizontally mounted furnace. The heating rate is maintained by manually increasing the Variac setting while the temperature and differential thermal effect are measured by reading deflections on a Rubicon 3400H galvanometer.

The two units produced curves which gave values to within ±5% of those obtained on our unit and which were also sensitive to a minimum of 5% of one component. Greater accuracy could be obtained by running a series of standard mixtures on each unit. It seems apparent that almost any type of apparatus may be used providing the following procedures are observed:

1. Not less than 0.6 ml. of sample should be used.
2. The sample should have the consistency of a thick paste.
3. A lid should cover the sample.
4. Rate of heating should approximate 10° C./min.

Summary

A differential thermal analysis technique is suggested for quantitative analysis of the relative amounts of endellite versus halloysite and or kaolinite in a sample. The endellite-ethylene glycol complex was found to have an endothermal peak at 500° C. as contrasted with the usual halloysite and kaolinite peak at about 575° C. A series of mixtures of endellite complex and halloysite and of endellite complex and kaolinite were run. A graphical solution is also presented utilizing ratios of the peak heights. The effect of four common impurities has been evaluated and the method has been tested on two other differential thermal units.

Acknowledgments

The authors wish to thank Dr. C. D. Jeffries of the Department of Agronomy and Mr. R. M. Gruver of the Division of Ceramics for running samples in their differential thermal analysis units, Dr. M. L. Keith of the Division of Geochemistry for providing reference quartz, Dr. C. E. Weaver, Division of Mineralogy for providing the mixed layer clay, Dr. T. S. Lovering of the U. S. Geological Survey for providing the endellite, and Mr. Walter Yahn for his assistance in the laboratory.

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A LATH SHAPED NON-EXPANDED DIOCTAHEDRAL 2:1 CLAY MINERAL

Charles Edward Weaver*

Abstract

A lath-shaped clay mineral has been found in the Oswego graywacke near State College, Pennsylvania. Electron micrographs, *x*-ray and differential thermal analysis patterns and chemical analysis indicate that the clay is a non-expanded dioctahedral 2:1 mineral with less iron and magnesium than the average illite. The clay is found both in the graywacke itself and in one to six inch beds of clay occurring within the formation.

Introduction

A clay mineral of the non-expanded dioctahedral 2:1 type (illite) showing a different morphology from that of previously studied representatives of this group has been found in samples of the Oswego formation of Upper Ordovician Age in the vicinity of State College, Pennsylvania. Because of the unusually good crystallographic development of the minute lath-shaped particles, a mineralogical study has been made of the clays associated with the Oswego and overlying Juniata formations to learn more of the nature and distribution of this clay mineral.

The Oswego formation is a graywacke consisting of approximately 800 feet of thick-bedded greenish-gray sandstone, siltstone and shale. It also contains numerous beds, one to six inches thick, which are 95 per cent clay. The Juniata graywacke is similar in composition but is largely red in color.

The following samples were collected at State College:

Sample	*Feet Below Juniata Shale*	*Description*
OS-1	200	Gray, six inch clay bed
-2	212	Gray, one inch clay bed
-3	227	Gray, one inch clay bed
-4	280	Gray, two inch clay bed
-5	296	Gray, two inch clay bed and clay gulls
-6	304	Gray, six inch clay bed
-7	370	Tan, moderately weathered shale
-8	203	Gray, highly weathered, fine grained graywacke
-9	300	Gray, highly weathered, medium grained graywacke
-10	300	Gray, fresh, medium grained graywacke

Several samples of the Juniata red shale (Ju-1, 2, 3) were collected about 100 feet above the Oswego contact.

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Additional samples of the Oswego clay beds were collected near Williamsport, 60 miles northeast of State College. Here it was noticed that the parallel breaks (commonly called bedding) in the coarse sandstone contain thin, one sixteenth to one quarter inch layers of clay similar to that in the larger beds.

MINERALOGY

Tuttle (1940), in his study of the Oswego and Juniata graywackes, showed that the average composition of the Oswego is 40 per cent rock fragments (slates, phyllites, argillites, schists, etc.), 50 per cent quartz grains, 10 per cent clay, and from less than one to fifteen per cent feldspar (albite, andesine, microcline). The Juniata is similar except for the presence of considerable iron oxide.

He found that the non-opaque heavy minerals are predominantly zircon and tourmaline which vary in shape from round to angular with apatite and chlorite being fairly abundant. Leucoxene, anatase, and iron oxides are the most common opaque heavy minerals.

In the present study it was found that the coarser grains which comprise approximately five per cent of the clay beds are similar to the grains in the sandstone. The heavy minerals from the shale and the clay beds are identical to those in the sandstone. The opaque minerals pyrite, leucoxene, and anatase, comprise about 90 per cent of the total heavy mineral concentrate. The non-opaques are largely sub-round zircon and tourmaline grains. (The zircon-tourmaline ratio varies from 1:3 in the clay beds and shale to 6:1 in the sandstone.)

ELECTRON MICROSCOPE DATA

The shape of the particles is the most interesting feature of the clay. To serve as a basis of comparison Fig. 1 is an electron micrograph of Fithian illite and Fig. 2 shows the non-expanded dioctahedral 2:1 clay mineral found in the Juniata formation. The morphology of the latter is typical of many illites, of brammallite, and of mixed-layer clays.

Figure 3 illustrates the appearance of clay particles in the less than one micron fraction of sample OS-8 from a sandstone member of the Oswego formation. Approximately half of the particles are sub-equant flakes while the remainder are thin, well-developed laths which range from 0.2 to 6 microns in length. The shadows cast by the metal shadowing technique show that some of the thinner laths are 40 to 60 Å thick indicating a stacking of 4 to 6 layers of 10 Å thickness.

Individual laths are the most common, but quite frequently they occur together in a parallel arrangement as equant or rectangular flakes (Figs.

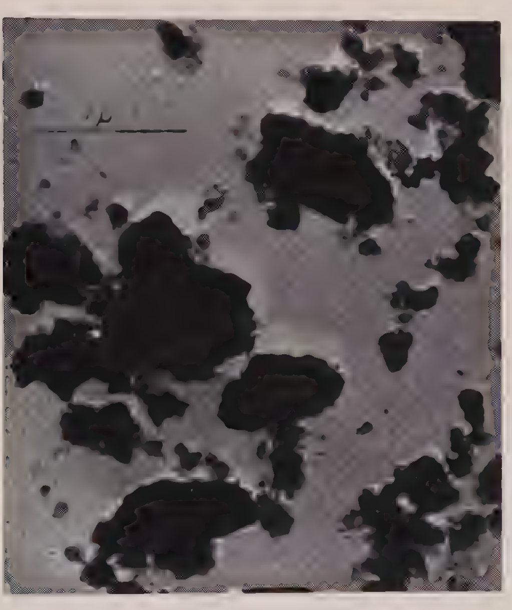

FIG. 1. Electron micrograph of the $<1\mu$ fraction of Fithian illite showing irregular small flakes. Mag. 27,600.

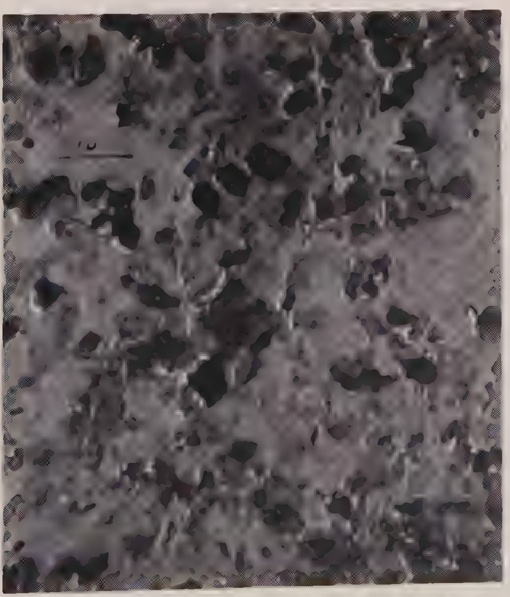

Fig. 2. Electron micrograph of the $<1\mu$ fraction of Juniata shale showing thin, subequant flakes with sharp boundaries typical of many illites. Mag. 13,500

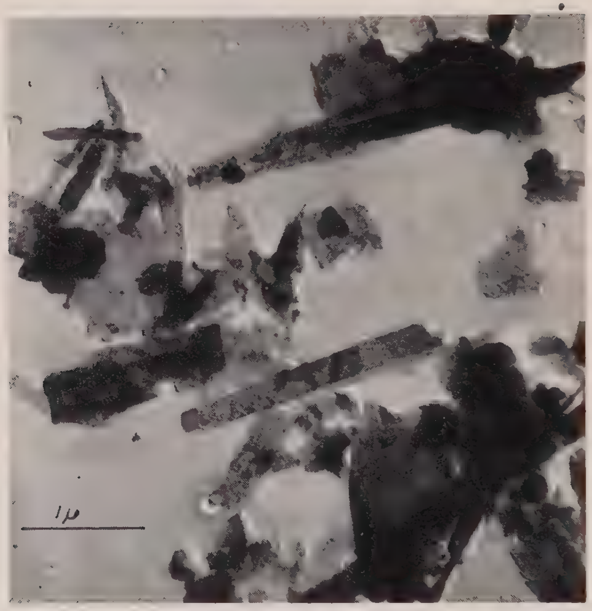

Fig. 3. Electron micrograph of the $<1\mu$ fraction of a sandstone member of the Oswego graywacke (OS-8) showing both sub-equant and lath shaped flakes. Mag. 21,400.

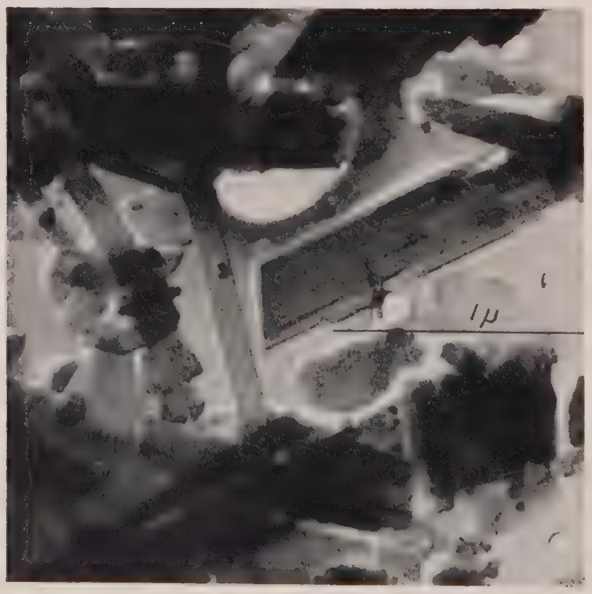

Fig. 4. Electron micrograph of sample OS-8 showing the step-like surface found on many large laths. Mag. 43,200.

4 and 5). From the evidence now available it is uncertain whether the laths result from cleavage of the flakes or by an oriented agglomeration of lath-shaped units. Figure 4 contains a flake which has a step-like surface. Figure 5 has several wide lath-shaped flakes with smooth surfaces which give no hint of being composed of parallel laths; however, the indentations on the ends of the flakes show that the larger laths consist of smaller laths in parallel orientation. It is possible that many of the smooth surfaced, sub-equant flakes are composed of, or would break into laths.

The smooth, sharp edge of the end of the lath seen in Fig. 4 was observed on a number of particles. The well-defined angles between sides and ends suggest that the laths often terminate in crystal faces.

The samples from the clay beds (OS-1 to -6) are composed largely of thin, angular, sub-equant flakes (Fig. 6). However, lath-shaped particles are frequently found. Figure 6 shows an aggregate of laths lying on a group of sub-equant flakes. The laths are similar in size and shape to those found in the clay fraction of the sandstone.

The electron micrographs of clay from the shale (OS-7) show that it consists largely of thin flakes many of which are sub-equant while others are elongated and have a lath-like shape. The laths are wider than those previously described and have a length-to-width ratio of about 4:1. In addition, the pictures show numerous well-developed halloysite tubes.

The clay from the Juniata shale samples consists entirely of thin, angular, sub-equant flakes.

X-RAY DATA

Powder photographs of the less than one micron fractions of both Oswego and Juniata samples indicate that all the clays are of the non-expanded dioctahedral 2:1 (illite) variety. Table 1 contains a list of the *d* values for a sample from a clay bed (OS-1) and for a sample of the less than one micron fraction of the sandstone (OS-8). These values compare quite well with those given by Grim, Bray, and Bradley (1937) for the Calhoun illite.

However, the Oswego and Juniata clays contain several lines (3.04, 1.94, 1.42, and 1.26 Å) which were not given by the Calhoun illite but do occur in muscovite patterns. Several of these values were reported in patterns of the Ballater illite (McKenzie, Walker, and Hart—1949) and Ordovician K-bentonite (Weaver –1952). The Juniata samples contain a small amount of chlorite.

The pattern of the sample containing few laths (OS-1) has two very weak lines, 3.19 and 2.98 Å, which do not occur in the patterns of the lath-shaped clay (OS-8).

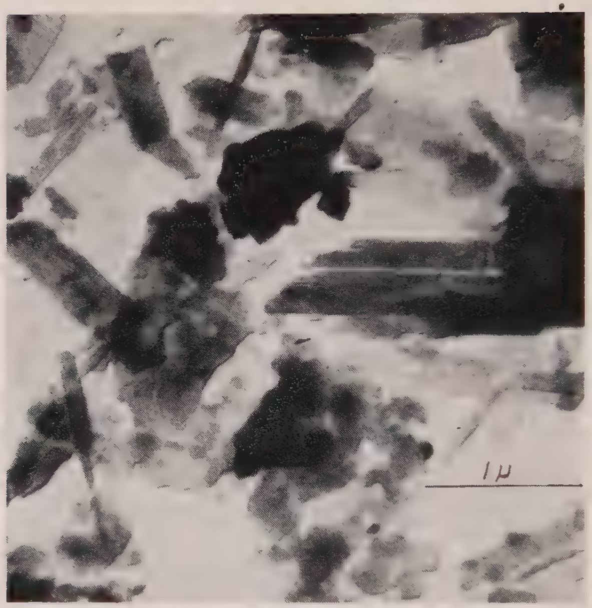

FIG. 5. Electron micrograph of sample OS-8 showing several wide lath-shaped flakes with smooth surfaces but with indentations on the end indicating that the larger laths are composed of, or may split into smaller ones. Mag. 27,200.

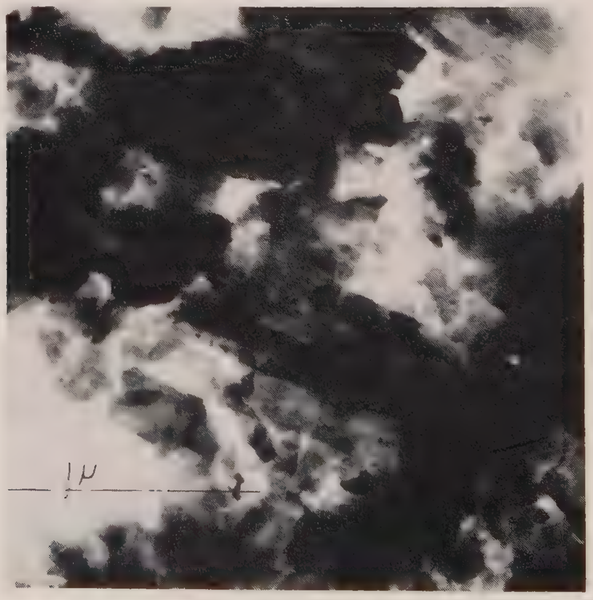

FIG. 6. Electron micrograph of the $<1\mu$ fraction of an Oswego clay bed (OS-4) showing sub-equant flakes overlain by a large aggregate of parallel laths. Mag. 43,200.

X-ray patterns of the lath-shaped material treated with ethylene glycol and with $CaCl_2$ show no indications of any expanded layers. *X*-ray spectrometer patterns of oriented slides of the lath-shaped clay show the same 001 periodicity and intensity as do the sub-equant flakes, indicating that the "*c*" axis is perpendicular or nearly perpendicular to the plane of the laths. The patterns of the oriented material also contain a broad, very weak peak at 7.2 Å possibly caused by a small amount of kaolinite.

TABLE 1

Indices	Calhoun Illite		OS-1		OS-8	
	d	*I*	*d*	*I*	*d*	*I*
002	9.98 Å	S	10.0 Å	S	10.0 Å	S
004	4.97	W	4.9	W	5.0	W
110	4.47	S	4.46	SS	4.45	S
022	4.11	WW	4.10	WW	4.10	WW
023	3.7	WW	3.65	W	3.66	W
$11\bar{4}$	3.4	WW	3.50	W	3.46	WW
006	3.31	M	3.32	M	3.32	M
114	3.2	WW	3.19	WW	—	
$11\bar{5}$	—		3.04	WW	3.05	W
025	2.98	W	2.98	WW	—	
115	2.84	WW	2.83	WW	2.82	WW
$20\bar{2}$	2.56	S	2.56	SS	2.56	SS
$13\bar{3}$	2.44	W	2.44	WW	2.44	WW
133	2.38	M	2.37	M	2.37	M
$22\bar{1}$	2.24	M	2.24	M	2.24	M
$22\bar{3}$	2.18	W	2.19	W	2.19	W
043	2.11	W	2.13	W	2.13	W
0010	1.98	M	1.98	S	1.98	S
206	—		1.94	WW	1.94	WW
$13\overline{10}$	1.65	W	1.65	W	1.65	M
312	1.64	M	—		—	
060	1.50	S	1.49	S	1.49	S
0014	—		1.42	WW	1.42	WW
335	1.34	WW	1.34	WW	1.34	WW
			1.33	W	1.33	W
400	1.29	M	1.29	WW	1.29	M
402	—		1.26	WW	1.26	WW
0016	1.24	W	1.24	WW	1.24	W

SS = very strong.
S = strong.
M = moderate.
W = weak.
WW = very weak.

CHEMICAL DATA

A chemical analysis of the less than one micron fraction of the sandstone sample is given in Table 2. *X*-ray data indicate that a small amount, less than five per cent, of kaolinite and two or three per cent of quartz are present. As the aggregate composition of these impurities would be approximately equal to that of a 2:1 dioctahedral clay the analysis was not recomputed.

TABLE 2

	Oswego (OS-8) Lath-shaped Clay	Average of Five Illites
SiO_2	49.67	48.95
Al_2O_3	27.31	25.03
Fe_2O_3	2.96	7.29
FeO	—	1.61
TiO_2	0.23	0.51
MgO	1.09	3.10
CaO	0.29	0.29
Na_2O	0.10	0.15
K_2O	7.26	6.03
MnO	0.06	—
H_2O	10.50	9.26
	99.47	100.22

Oswego: Analyzed by G. Kunze, Department of Agronomy, The Pennsylvania State College.

Illite: Average of five analyses, Grim *et al.* (1937).

The only significant difference between the Oswego sample and the average of the five illites is in the amount of iron and magnesium present; the five illites contain approximately three times as much of both as does the Oswego material.

Neglecting the TiO_2 and MnO the structural formula as computed by the method of Ross and Hendricks (1945) is:

$$K_{.66}\frac{Ca}{2}_{.04}Na_{.01}$$

$$[Al_{1.71}Fe'''_{.16}Mg_{.11}]\quad[Al_{.53}Si_{3.47}]\quad O_{10}[OH]_2.$$

As compared to other illites, the amount of replacement of Si by Al in the tetrahedral layer (13%) is similar, whereas that of Al by Fe and Mg in the octahedral layer (13%) is much less.

DIFFERENTIAL THERMAL DATA

Differential thermal analysis curves of the less than one micron fraction of a number of samples are shown in Fig. 7. All curves show a weak peak near 110° caused by the hygroscopic moisture. Curves 1 and 2, from the Oswego clay beds have a broad hydroxyl-endothermal peak with an apex at 625° C. Curve 4 of the clay from the Juniata shale is quite similar. Curve 3 of the Oswego clay from Williamsport has a broad peak similar to the others but in addition has a small dip at 575° C. possibly caused by a small amount of kaolinite. Curve 5 is typical of the Oswego shale. This curve shows the broad hydroxyl-endothermal peak of the illite but in addition has a sizable peak at 580° C., extending below the broad peak, which is probably caused by the halloysite that was noted in the electron micrographs. Curve 6 was obtained from the lath-shaped clay from a sandstone sample (OS-8). The curve has the same type of broad hydroxylendothermal peak as the others but has apices at both 550° C. and 600° C., one of which may be caused by a small amount of kaolinite. The exothermal peak at 475° C. may be caused by pyrite or organic material.

The hydroxyl-endothermal peak in these clays is higher than those found in most illites. This is possibly due to the small amount of Fe present.

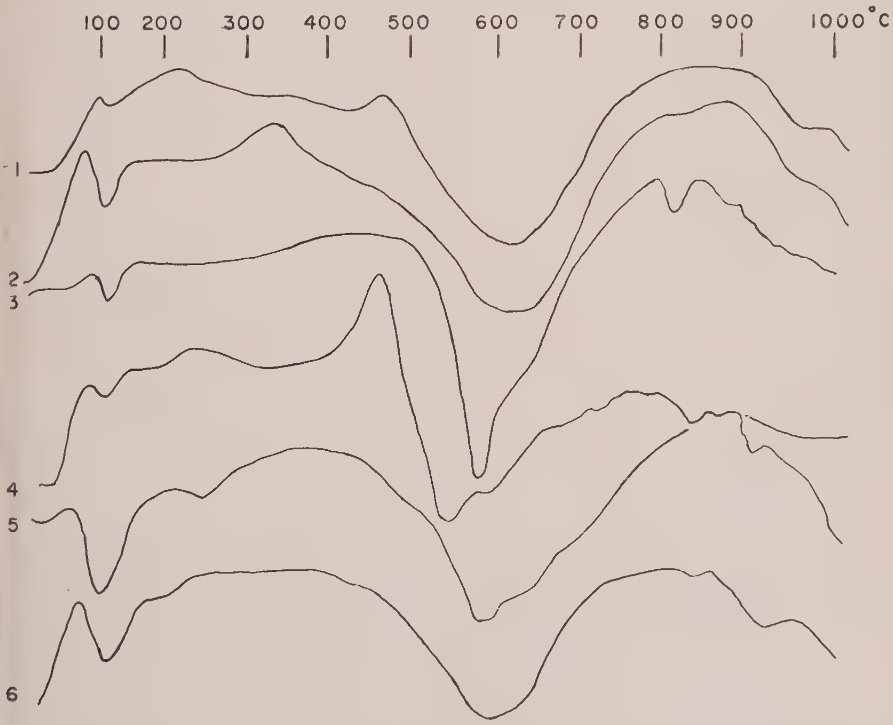

FIG. 7. Differential Thermal Curves: (1) Oswego clay bed, OS-1; (2) Oswego clay bed, OS-4; (3) Oswego shale; (4) Oswego graywacke, OS-8; (5) Oswego clay bed from Williamsport, Pa.; (6) Juniata shale.

ORIGIN OF LATHS

In the 2:1 clay minerals predominant growth along one axis is usually considered to be caused by a directional strain in the mineral lattice due to the inequality in size between the octahedral layer and the two tetrahedral layers. Whereas a similar misfit in the 1:1 lattice causes curved plates and in many cases tubes, here the misfit layer is sandwiched on both sides by layers which prevent curling of the sheets but do not relieve the strain. The inequality in size is usually caused by the presence of ions larger than Al, usually Mg and Fe‴, i.e. hectorite and nontronite, (Ross and Hendricks—1945). However, in the Oswego material Fe‴ and Mg are both present in amounts considerably less than is generally found in other non-expanded dioctahedral 2:1 clays.

The sharp angles found on many of the laths plus the small amount of isomorphous replacement in the octrahedral layer suggest that this may be a more highly crystallized form of a non-expanded dioctahedral 2:1 mineral and have less strain than clay particles of a similar type described in the past.

ORIGIN OF CLAY

The examination of the heavy and light minerals in the clay beds revealed no trace of the volcanic minerals which are common in many Ordovician clay beds. The heavy minerals are similar to those of the coarser fractions of the Oswego and Juniata graywackes. It is concluded, therefore, that the clay had the same source area as the coarser material and is detrital in origin.

The concentration of the clay into relatively pure layers one to six inches thick is possibly caused by a winnowing of the clay from the coarser sand by wave action. Ocean currents would carry the clay away in suspension until calmer water was reached and it would then be deposited for varying lengths of time until the current became strong enough to include silt and sand grains in its load. The thin clay layers between the bedding planes were probably deposited by a similar process during a much shorter period of time.

The lath-shaped clay, which was not found in the Juniata, is most abundant in the Oswego sandstone samples. As it is also present, but in lesser amounts, in the relatively pure clay beds it is possible that this clay is detrital; however, within the sandstone samples it appears to be more abundant in the more highly weathered samples suggesting that it may be secondary, possibly from the weathering of the feldspar.

ACKNOWLEDGMENTS

The writer is indebted to Dr. Thomas F. Bates for his help and criticism in preparing this paper. Thanks are given to Dr. George Kunze for the chemical analysis and to Mr. Joseph Comer for the electron micrographs.

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PRESENTATION OF THE ROEBLING MEDAL TO FREDERICK EUGENE WRIGHT

Edward H. Kraus, *University of Michigan, Ann Arbor, Michigan.*

Twenty-three years ago, in 1929, I suggested at the annual meeting of the Society that awards, either medals or prizes, be established to recognize noteworthy achievement in the sciences represented by the Society. It was not until 1936 that the Council authorized the designing of a medal and voted that it be called the Washington A. Roebling Medal. The first award was made in 1937 to Charles Palache. Thus far, the medal has been awarded ten times. Our universities are represented by three recipients, the United States Geological Survey and the Geophysical Laboratory by two each, Great Britain by two, and Switzerland by one. Today the medal is to be awarded to one who for more than a half century has been an active contributor to the advancement of the sciences represented by our Society. He has also been active in related fields and has long had an international reputation as a distinguished scientist.

Frederick Eugene Wright was born at Marquette, Michigan, October 16, 1877, where his father was stationed as state geologist. The elder Wright was well trained in mineralogy and geology for he had studied these subjects in Germany and Sweden. He was one of the first in this country to prepare thin sections of rocks. Unfortunately he died at an early age in 1888. Shortly after his death Mrs. Wright with her three sons moved to Ann Arbor, Michigan. Fred, as he is affectionately known, attended the public schools and was graduated from the Ann Arbor High School in 1895. The family then moved to Germany where Fred, the oldest of the three boys, was a student at the Realgymnasium at Weimar for one year. In 1896 he enrolled at the University of Heidelberg where, following in the footsteps of his father, he pursued intensive studies in mineralogy, petrology, and geology, as well as in chemistry, physics, and mathematics under the direction of such distinguished scientists as Harry Rosenbusch, Victor Goldschmidt, Wilhelm Salomon, Adolph Sauer, Viktor Meyer, P. Lenard, G. Quincke, L. Koenigsberger, and others. In December 1900 the degree of doctor of philosophy (summa cum laude) was awarded to him at the age of 23. It is of interest to note that while in Heidelberg Dr. Wright spent a period in the shop of the well-known instrument maker Peter Stöe. This experience later proved very helpful in developing the various optical instruments and accessories with which the name of F. E. Wright has long been associated.

After his return to the United States Dr. Wright served as instructor in petrology at the Michigan College of Mines (now the Michigan Institute of Mining and Technology) at Houghton, from 1901 to 1904, and for one year he was also assistant geologist on the Michigan Survey. He then became associated with the United States Geological Survey. In 1906 he was appointed as petrologist to the Geophysical Laboratory, which had recently been established, where he served until retirement in 1944. He was one of the early members of the staff of the laboratory.

Dr. Wright's bibliography includes over 140 papers dealing directly with mineralogical and petrological subjects. Time does not permit a detailed review of these papers; suffice it to say that they include the classical articles on Quartz as a Geologic Thermometer with E. S. Larsen, the designing of an improved petrographic microscope and of many testplates and accessories, as well as numerous studies on the optical properties of various minerals and synthetic products. The publication in 1911 of his authoritative *Methods of Petrographic-Microscopic Research: Their Accuracy and Range of Application* did much to stimulate the use of the petrographic microscope in this country. Indeed the present wide use of the petrographic microscope in research not only in our sciences but also in related fields, is due in large measure to Dr. Wright's early and persistent advocacy of its value. He was a pioneer in this country in crystal optics and the application of polarizing microscopic methods.

In addition to this long list of papers, Dr. Wright made important contributions during World War I to the development and manufacture of optical glass in this country, which previously had to be imported from Europe. These studies were sponsored by the Ordnance Department and resulted in the publication in 1921 of the book entitled *The Manufacture of Optical Glass and of Optical Systems*. For these services he attained the rank of Colonel in the Ordnance Reserve Corps. Due to his expert knowledge in this field he served during World War II as Civilian Adviser to the Joint Optics Committee of the Army and Navy Munitions Board and received from the Army the *Gold Medal for Exceptional Civilian Services*.

Reference must also be made to Dr. Wright's researches and activities as Chairman of the Committee on the Study of the Surface Features of the Moon. These studies, which he began in 1924 and continued until 1939, were made at the Mt. Wilson Observatory. The results are recorded in the annual reports of the Committee in the Year Book of the Carnegie Institution of Washington.

Dr. Wright is a member or fellow of many scientific societies in this country and abroad, which include the National Academy of Sciences, the Mineralogical and Optical Societies of America of which he is a past president, and the Geological Society of America. In 1941 the honorary degree of Doctor of Science was conferred upon him by the University of Michigan.

Mr. President: It is indeed a high privilege to present Frederick Eugene Wright as the recipient of the Washington A. Roebling Medal. In making this award both Dr. Wright and the Society are being honored.

ACCEPTANCE OF THE ROEBLING MEDAL OF THE MINERALOGICAL SOCIETY OF AMERICA

Fred E. Wright, *2134 Wyoming Ave., N. W., Washington 8, D. C.*

It gives me real pleasure to receive the Roebling Medal of the Mineralogical Society of America and to realize that award was made on the basis of the recommendation by its Council.

You have heard from Doctor Kraus that, many years ago, he proposed to the Mineralogical Society that it should establish a Medal to be awarded in recognition of work done in the field of minerals. Our Society owes much to him for his interest and participation in its affairs over the years; as one of the recipients of the Medal, I am grateful to him for his proposal. In his Presentation Address he mentioned many details of my own efforts; I marvel at his aptitude for digging out events in my life that had long since passed down memory's flow and been forgotten. This is in keeping with his reputation, the world over, for accuracy of statement and for reliability.

For me it is a special honor to receive a Medal named after the builders of the Brooklyn Suspension Bridge. Both father and son were remarkable men and civil engineers of the highest grade. The father, John A. Roebling, died in 1869; he had then received the contract to build the bridge; but actual construction had not been started at the time of his death which was the result of a serious injury incurred during a survey of the work to be done. He had designed the bridge; the drawings had been completed as far as was possible at that stage. It fell to the son to build the bridge. This he did, although he became seriously ill with caisson disease (bends) contracted in the course of his work. He had, in consequence, to direct the construction from his bedroom. In this effort he was aided by his wife, Emily Warren Roebling, who served as liaison between him and his engineering staff. The bridge has stood the test of time. It has recently been modified, to meet modern demands, by D. B. Steinman, a distinguished builder of bridges, whose book on the Roebling family and its accomplishments has been an inspiration to many of us. It was the son, Washington A. Roebling, who constructed the bridge and who gave a sum of money to the Mineralogical Society; part of this

FRED. E. WRIGHT

Recipient of the Roebling Medal of the Mineralogical Society of America fund is devoted to the Medal Awards. The son died in 1926; he was an enthusiastic collector of minerals; his wonderfully complete mineral collection was given, after his death, by his son, John, grandson of John A. Roebling, to the National Museum of the Smithsonian Institution at Washington together with a large maintenance fund.

It is customary for the recipient of the Roebling Medal to refer briefly to a phase of his own work. I have chosen to discuss some of the results of our investigation of the materials exposed at the moon's surface. To one trained in geology and mineralogy, the first essential to progress is knowledge of the materials in the field and of their behavior toward agents that tend to modify them. Thus he asks: "What are the materials at the moon's surface?" and "what is the physical and chemical behavior of these materials under lunar surface conditions?"

Unfortunately we cannot visit the moon and learn at first hand about its rocks and their environment. The best we can do is to study the reflected and scattered sun's rays (moonbeams) and note the effects that lunar surface materials have on them; comparison of these effects with those produced by terrestrial materials may enable us to form an idea of their composition. The effects are of two kinds: (1) Changes in the relative intensity of light from the ultraviolet into the infrared; also the changes in the total intensity of light scattered at various angles of incidence; (2) Degree of polarization of the scattered rays for various angles between incident and reflected or scattered rays.

With the aid of a special eyepiece for measuring the amount of plane polarization in a beam of light, the percentage of plane polarization in the light from twenty-four selected points on the moon's surface was determined for various phases of the moon during eleven lunations. The relative accuracy of these measures is less than one-fifth of one per cent, plus or minus.

The problem is analogous to that of the determination of minerals under the petrographic microscope. In the one case, the changes produced in light on diffuse reflection and scattering are analyzed; in the second case, those resulting from transmission through mineral sections and grains are determined. With lunar materials the changes are unfortunately few and cannot be ascertained with the degree of accuracy possible in mineral plates. The observer has no choice but to infer the nature of its surface materials from the slight changes they produce in sun's rays on reflection and scattering. If satisfactory conclusions are to be drawn, the measurements should be made with high accuracy by independent methods and through complete lunations.

The phenomena observed are chiefly those of scattered (diffracted) light combined with absorption and internal and external reflections;

under these conditions the character of the surfaces which scatter the light plays an extremely important role and gives rise to phenomena that reflection and refraction alone do not explain. When the angle between the incident and the radiated beam is small, between 10° and 30° depending on the substance and on its surface, the amount of polarized light in the scattered is zero. For still smaller angles the scattered beam is again polarized, but the plane of its vibration is in the plane of incidence, rather than normal thereto.

Lyot, the first to observe this phenomenon in moonlight, called it negative polarization. It appears about two days before full moon and continues for an equal period after full moon. Terrestrial materials exhibit the same phenomenon and about the same variation. The percentage amount of negative polarization rarely exceeds one per cent. In terrestrial materials it varies with the substance and with the scattering surface. As a diagnostic feature negative polarization has been of little value; it is an important element in the theory of the scattering of light and is being investigated further.

The visual measurements have been made with an eyepiece which consists of a tilting-plate compensator together with a detector for ascertaining the exact point of compensation. The field of the eyepiece is a divided photometric field in which two factors, namely, equality of illumination and exact alignment of Savart fringes, are the criteria used in making a measurement; it is the combination of these two factors that renders the method so accurate. With this eyepiece the percentage amount of polarization can be determined to 0.2 per cent. The plate of the compensator can be tilted to 70°; for this angle of tilt the percentage amount of plane polarization introduced by a single glass plate or film of celluloid of refractive index, 1.505, is about 26 per cent; at a 60° angle of tilt it is 18 per cent; with a compensator of two plates mounted in parallel and tilted at 60°, the percentage amount is increased to 30 per cent; with 4 plates in parallel, to 46 per cent; with 6 parallel plates, to 56 per cent; with 12 parallel plates, to 71 per cent.

For observations on the moon the single-plate compensator suffices; the maximal polarization in moonlight from the dark areas or Mare is approximately 17 per cent; from the bright lunar mountain areas it rarely exceeds 8 per cent. For observations on terrestrial materials more than one tilting plate is needed to compensate the much higher amounts, exceeding 60 per cent, of plane polarization produced by dark colored rocks, such as basalts and peridotites.

With the polarization eyepiece, equipped with tilting compensators of one, two, and six parallel plates, polarization measurements have been made on fifty selected terrestrial substances illuminated by sunlight incident at different angles and viewed from different directions.

The results of visual measurements of polarization in light scattered by the moon's surface and by terrestrial materials have been plotted on graphs and indicate that basalts and other rocks low in silica are not exposed on the surface of the moon. To judge from the results the rocks are in the form of pumice and range in composition from rhyolites high in silica to trachytes and andesites lower in silica, but above 55 per cent. The surmise that a thin layer of dust due to migration of alumina and other components of the underlying substances to the surface and to atom and molecule disintegration through bombardment by electrons and neutrons emanating from the sun weakens this conclusion which rests solely on polarization measurements. The results do not prove the existence of silicate rocks at the surface. Silica and the silicates show metallic reflection in the region of 8 to 10 mμ in the infrared; but the intensity of solar radiation in that region is so low that Pettit and Nicholson at Mt. Wilson were unable to detect it in the presence of planetary heat, namely heat absorbed at short wave lengths and radiated at much longer wave lengths. The rate of cooling during a lunar eclipse, as measured by Pettit, proved that an excellent heat insulating material covers the lunar surface, such that during a lunar eclipse the fall in surface temperature approaches several hundred degrees centigrade. If the material is finely divided dust, it in turn is fairly transparent to rays in the visible spectrum; were the layer dark, it would show a much greater percentage of polarization than is now the case. It has long been known that the Lambert cosine law of scattering of light does not hold accurately, thus indicating the presence of pumiceous material. Furthermore as result of a recent work on microwave observations, Jaeger and Harper and others have shown that there is possibly a thin crust of dust only a few millimeters thick covering the pumice. In no case has evidence been found of solid rock, such as granite, syenite, or diorite, outcropping at the surface. Even the inner slopes of craters give no clue to the presence of solid crystalline rock.

This work is fascinating in part because of the distance at which the lunar surface materials have to be studied; this factor alone adds much to the difficulties of the problem; in part because of our limited knowledge of the behavior of the materials exposed to conditions existing at the lunar surface. To us it is a strange world in which our background of terrestrial experience avails but little and may even be a hindrance.

Permit me again to emphasize the gratitude and pleasure at the award to me of the Roebling Medal. It will ever be an inspiration to me to be reminded of the courage, the tenacity, and the thoroughness of the giver, Washington A. Roebling.

PRESENTATION OF THE MINERALOGICAL SOCIETY OF AMERICA AWARD TO FREDERICK HENRY STEWART*

WALDEMAR T. SCHALLER, *U. S. Geological Survey, Washington, D. C.*

Mr. President, Ladies and Gentlemen:

Why do any of us spend our lives with minerals and aggregates of minerals—the rocks? It must be that consuming interest in things mineralogic, an interest that makes us study their properties, and makes us want to know why minerals and rocks are what they are. To get an answer, be it right or wrong, thrills us no end. This attainment of an answer makes us feel that it has been well worthwhile to do what we have done.

Minerals are but Nature's Chemicals and so we are plunged at once into all the intricacies of chemistry, and our field of endeavor widens greatly, so greatly in fact, that we soon find that none of us can do everything all by himself. Then too, we are human beings. We don't want to be ourselves alone. We are gregarious and like to be with and to work with, other human beings, to share our discoveries, such as they are, with others, and to hear their comments, praiseworthy or otherwise. Being gregarious, we congregate together and form societies and establish means of conveying our findings to those who, because of their distant locations, cannot always be close to us.

We bask gleefully in our discoveries. We get an even greater thrill when others say "well done." Numerous awards have been established by organized societies to honor those whose life work was "well done." But what of those in early life who, though they have already accomplished many things, are yet but in their beginning? It is in line with this thought, that Bowen, last year, aptly described the new Award of the Mineralogical Society of America as "inspired"—the Award "to be given for a specific accomplishment indicating great promise and to serve as a spur to the fulfillment of that promise by one still vigorous enough to respond."

That the second recipient of the Award, Dr. F. H. Stewart of Durham University, England, completely meets these specifications all of you who have had the privilege of being with him these last few weeks, will agree fully. As he wrote me last May: "The decision of The Mineralogical Society of America to give their Award to me came as a complete and delightful surprise, and I feel most humbly grateful to be thus honoured. I can assure you that this will act as a most powerful incentive to me in my further research."

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* Stewart, Dr. Frederick Henry, Department of Geology, University of Durham, England. Born in Aberdeen, Scotland, January 16, 1916. B.Sc., with honours in geology, Aberdeen University, 1937; Post-graduate research under Prof. T. C. Phemister, Aberdeen University, 1937–1939; under Prof. C. E. Tilley, Cambridge University, 1939–1941; Ph.D., Cambridge, 1941. Physico-Chemical Section of Research Department, Imperial Chemical Industries, 1941–1943. Department of Geology, Durham University, Durham, England, since 1943. Received Lyell Fund Award, Geological Society of London, 1951.

The Award was made to Dr. Stewart for his investigation of the English evaporites of Permian age, saline deposits showing many similarities to our own Permian saline deposits in New Mexico and Texas. His mineralogic and petrologic study of the English evaporites, given in three papers covering a hundred printed pages, was so thorough and of such excellent quality, that not only are they an "outstanding contribution within the fields" of our Society but they also show great promise of work—"well done"—yet to come.

I take great pleasure in presenting to you, Dr. Frederick H. Stewart of the Department of Geology, University of Durham, England, as the second recipient of the Award of the Mineralogical Society of America.

ACCEPTANCE OF THE MINERALOGICAL SOCIETY OF AMERICA AWARD

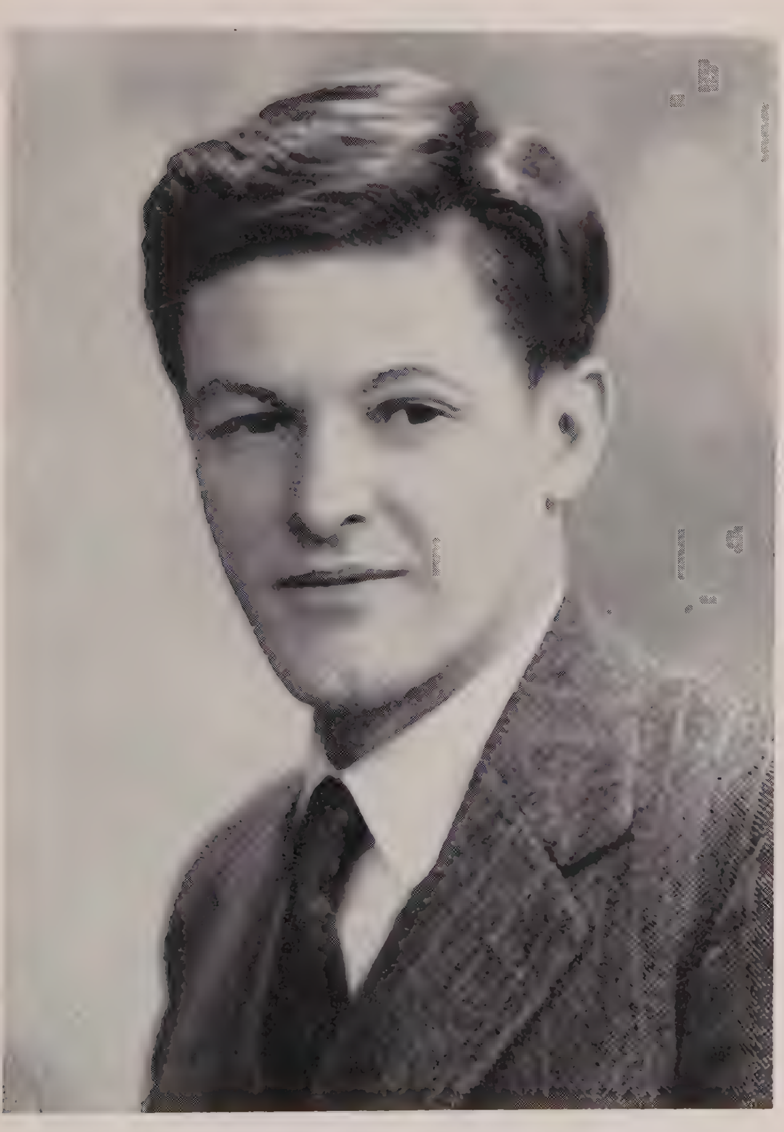
Frederick H. Stewart, *Department of Geology, University of Durham, Durham, England.*

Mr. President, Dr. Schaller, Fellows and Members of the Mineralogical Society of America:

It is impossible for me to convey adequately my feelings of gratitude, pleasure, and surprise when I received a letter from your President saying that this Award was to be bestowed on me. I can only say that I am deeply sensible of the honour which this Society has done me, an honour which will certainly act as a great incentive to me in my future work. It is particularly gratifying to be put in the company of a man of such attainments as last year's recipient, Dr. Tuttle.

This honour should really belong to my teachers and colleagues: my uncle, Dr. William Alexander, and the late Professor W. T. Gordon, who first aroused in me an intense interest in minerals and rocks at a very early age; my teachers at Aberdeen and Cambridge Universities, to whom I owe more than I can hope to repay; my colleagues at Durham University, who have given me every help and encouragement; these should share the credit of this Award.

Dr. Schaller has been very kind, and I am especially glad that my work should have been favoured by him, because I have followed in the



FREDERICK HENRY STEWART

Recipient of the Mineralogical Society of America Award footsteps of his classical work on the petrology of the Texas-New Mexico evaporites. Without his work as a secure foundation I should never have gone anything like as far as I have.

Since one of the primary objects of this Award is that it should act as an incentive towards further work, it is appropriate that I should say a few words about my present work. With Professor L. R. Wager, I am engaged in a long-term study of the Tertiary igneous rocks of the island of Skye. I am also continuing my evaporite work with a study of the material from the recent exploratory boreholes of Fisons Limited in the British potash field. On the argumentative question of primary gypsum versus primary anhydrite in marine evaporites, I have now obtained evidence suggesting that much of the $CaSO_4$ was originally deposited as gypsum in the British Permian at least. The results of this work are in the hands of the printers. To my mind the original character of the lower zones of these deposits can now be brought much more nearly into line with the theoretical succession in marine evaporites. Mineralogical and petrographic work is bringing to light extremely complex replacement phenomena in the upper zones. To explain these, even Sherlock Holmes would have been hard put to it. As a simple petrological policeman I am still looking for places to put my large feet. However, I can assure you that I am looking as hard as I can, and that this Award will ensure that I continue to do so. As Dr. Bowen implied on the equivalent occasion last year, the bi-monthly receipt of the *American Mineralogist* will make my conscience work overtime at least six times a year.

I am particularly glad that I have been able to come here and receive this Award in person, and, while in America, to have been able to see some of your famous Universities and Institutions. The kindness which I have received has almost overwhelmed me, and I take this opportunity of expressing my most sincere thanks to all concerned, and particularly to your Secretary, Professor Hurlbut, who, by arranging a lecture tour for me, has made it possible for me to be here. In about ten days' time I shall see the famous evaporites of New Mexico, and that visit, together with the most stimulating discussions which I have had with Dr. Schaller and many others, should make me far more capable than I was of fulfilling my responsibilities as a recipient of this Award.

Again, I thank the Society, in all humility, for this honour, and assure you that I shall do my utmost to be worthy of it.

MEMORIAL OF SAMUEL GEORGE GORDON

William Parrish, *Philips Laboratories, Inc., Irvington-on-Hudson, N. Y.*

The unexpected death of Samuel George Gordon came as a shock to his many personal friends and colleagues. The mineralogical world has lost one of its outstanding men. "Sam" was admired and respected by many people in all walks of life. He had a keen mind, was extremely competent, unassuming and was loyal to his friends.

He was born in Philadelphia, Pa., June 21, 1897, the son of Isadore and Flora Gordon. He is survived by his wife, Bertha May Landrum of Nacogdoches, Texas, and two sisters Julia and Gertrude; there were

Samuel George Gordon

1897–1952

no children. He died of a coronary occlusion on May 17, 1952, in Cincinnati, Ohio.

Gordon was one of the founders and a fellow of the Mineralogical Society of America. He played an important role in starting the *American Mineralogist* and was an associate editor from its beginning in 1916 until 1949. He was an early member of the Philadelphia Mineralogical Society, an officer-elect of the Crystallographic Society of America, American Crystallographic Association and the Mineralogical Society of Great Britain.

As a young boy in the years around 1911–12, Gordon took a course in mineralogy and geology under Professor Edgar T. Wherry at Wagner Free Institute of Science of Philadelphia. He and a number of other students who attended these courses joined the Philadelphia Mineralogical Society and formed the enthusiastic kernel of members who were mainly responsible for the continued success of this club. Gordon planned and led numerous collecting trips around the Philadelphia area and reports of the trips and exhibits of the specimens collected were for many years features of the Philadelphia Mineralogical Society's monthly meetings.

Gordon never had an opportunity for a complete formal university training. In addition to the lectures at Wagner he later took occasional courses at University of Pennsylvania and Drexel Institute and spent six months (1926 27) with Victor Goldschmidt at Heidelberg, Germany.

Gordon's long association with the Academy of Natural Sciences of Philadelphia began in 1913 when he became Jessup student and later assistant curator of minerals, working under Frank J. Keeley, curator of the Vaux collection. His many mineral collecting trips around the Philadelphia and adjoining areas gave the Academy the best existing collection of local minerals. When he was not yet 24 he completed *The Mineralogy of Pennsylvania* which became the standard source book on mineral localities in this state.

Gordon made five major mineralogical expeditions under Academy auspices to secure new specimens for the Vaux collection and to obtain new research material. On each of these trips he returned with 25 to 30 cases of minerals. On his first trip in 1921 he visited the famous mineral localities in the Andes of Peru, Bolivia and Chile, and collected magnificent suites of brilliant, highly modified tetrahedrite, bournonite, enargite, translucent cassiterite, many rarer and some new species. In 1923 he went to the Julianehaab District of Greenland and collected suites of aegirite and albite as well as kalithomsonite, rinkite, etc. In 1925 he returned to Bolivia and Chile, in 1929–30 he went to the Bolivian Andes and Africa (radioactive minerals, azurite and malachite, meteorites, etc.) and in 1938 he visited the Atacama Desert of Northern Chile. On the latter trip he took along a special three-wheel motorcycle provided with storage compartments for tools and specimens. This was characteristic of Gordon's disregard of personal comfort in order to minimize the Academy's expenses and to reach localities which otherwise would be inaccessible. His accounts of these trips in the Academy Year Book make interesting reading even today.

Gordon was particularly interested in the minerals of the great tin mines of Cerro de Llallagua, Bolivia. He visited these mines on three of his trips and examined and collected specimens from more than 100 working stopes as well as many of the accessible pillars in the abandoned stopes. His memoir, *The Mineralogy of the Tin Mines of Cerro de Llallagua, Bolivia*, published in 1944, contains some of the most remarkable crystal drawings ever made.

Gordon described nine new mineral species, shown in the following table.

Year	*Mineral*	*Composition*	*System*
1922	Vauxite	$FeAl_2(PO_4)_2(OH)_2 \cdot 7H_2O$	Triclinic
1922	Paravauxite	$FeAl_2(PO_4)_2(OH)_2 \cdot 8H_2O$	Triclinic
1926	Penroseite	$(Ni,Cu,Pb)Se_2$	Isometric
1926	Trudellite	$Al_{10}Cl_{12}(OH)_{12}(SO_4)_3 \cdot 30H_2O$?	Hexagonal?
1927	Metavauxite	$FeAl_2(PO_4)_2(OH)_2 \cdot 8H_2O$	Monoclinic
1941	Cadwaladerite	$Al(OH)_2Cl \cdot 4H_2O$	Amorphous
1941[a]	Sarmientite	$Fe_2(AsO_4)(SO_4)(OH) \cdot 5H_2O$	Monoclinic
1948[a]	Sanmartinite	$(Zn, Fe, Ca)WO_4$	Monoclinic
1950[b]	Wherryite	$Pb_4Cu(CO_3)(SO_4)_2(OH, Cl)_2O$?	(Biaxial)

[a] With Victorio Angelelli.

[b] With Joseph J. Fahey and E. B. Daggett.

Keeleyite, described in 1922, was later shown to be identical with zinkenite.

In 1930 Larsen and Shannon[1] gave the name *gordonite* to a new mineral, $MgAl_2(PO_4)_2(OH)_2 \cdot 8H_2O$, found in the phosphate nodules near Fairfield, Utah, after Gordon. It is related to paravauxite with Mg replacing Fe.

In 1928 Goldschmidt and Gordon published *Crystallographic Tables for the Determination of Minerals*. This book contains data on 759 minerals arranged in tabular form for identification with the two-circle goniometer. Gordon developed remarkable ability in using this classical technique to its fullest advantage. He could set up tiny faced mineral crystals and usually could make an identification faster than one could prepare an x-ray powder photograph. Gordon was extraordinary in the identification of minerals in hand specimens using only a hand magnifier and in micromounts using a binocular microscope. The latter is particularly difficult since the morphological habit of these tiny crystals is usually markedly different from those of larger crystals.

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[1] Larsen, Esper S., and Shannon, Earl V. The minerals of the phosphate nodules from near Fairfield, Utah: *Am. Mineral.*, **15**, 307–337 (1930).

Gordon was an expert in museum displays. The mineral and gem exhibits he prepared attracted the layman as well as the serious student. Many members of the Philadelphia Mineralogical Society were first attracted to mineralogy by his beautiful arrangements of the mineral collections. There is probably no mineral collection in the world that was displayed more efficiently and effectively. His magnificient exhibit of fluorescent minerals installed in 1928 has continued to attract thousands of visitors. He helped in the development of new types of museum cases which were so well designed the exhibit seemed to be floating in air.

Since the Academy is a public institution many people brought mineral and rock specimens for identification. Gordon spent a great deal of time doing this somewhat thankless chore and whenever he found a person with unusual interests in minerals he spent considerable time explaining the nature of the specimens and sometimes in encouraging the person to further studies.

In early 1942 Gordon obtained a leave of absence from the Academy to take a position in Office of the Chief Signal Officer, Washington, D. C. He joined our group (Quartz Crystal Coordination Section) working on the problems arising from the sudden unprecedented demand for quartz oscillator-plates. These plates which control the transmitting and receiving frequencies in communication equipment were in critically short supply for tanks, planes, walkie-talkies, etc. The manufacturing equipment and know-how had to be built very quickly and here Gordon showed some rare gifts. Since none of us knew the "best" orientation and cutting schemes at that time, we visited many plants to survey the problem. Sam always showed unusual perception in piecing together the procedures used from the noncrystallographic crude descriptions given by the workers using the methods. He quickly saw the importance of the parallelogram light figure obtained from etched XZ sections, the necessity for making an inspection after each step in the procedure and was a key figure in the development of procedures for cutting crystals which had no faces. He developed various rule-of-thumb procedures for quickly evaluating the efficiency of a plant, which were valuable in predicting production rates. His visits to the many crystal plants were always accompanied by amusing but accurate reports and personal letters which were so clear the reader felt that he was also there. In early 1943 we collaborated in the publication of *Manual for the Manufacture of Quartz Oscillator-Blanks*, the first integrated scheme worked out for large scale production of precision cut blanks. The *Symposium on Quartz Oscillator-Plates*, the May–June 1945 issue of the *American Mineralogist*, contains four papers by Gordon on this subject. He returned to the Academy in 1944 after the emergency had been successfully met.

Unfortunately the facilities and funds available at the Academy were never adequate to support more than the most modest work. Although Gordon's salary was far less than he could obtain as a professional mineralogist he was so devoted to his work that I never once heard him complain. His equipment consisted only of a two-circle reflecting goniometer, an old petrographic microscope and a few odd mineralogical tools. In recent years a war surplus *x*-ray machine was obtained and Gordon with the help of a friend constructed some crude cameras. Finally even this modest program could no longer be supported and Gordon reluctantly left the Academy in the fall of 1949.

He returned to Government work and became senior metallurgist on the NEPA Division of Fairchild Engine and Airplane Corporation at Oak Ridge. Here he conducted crystallographic studies of intermetallic compounds and high temperature alloys. He developed efficient schemes of using minute amounts of powder for *x*-ray analysis. When the project terminated in 1951 he was given glowing letters of appreciation for essential services. He then became a staff member of the Chemistry and Metallurgy Research Division at Los Alamos Scientific Laboratory and did *x*-ray work on classified problems. He remained there nearly a year and was hired to join the staff of the Air Nuclear Propulsion Division of General Electric Company, Lockland, Ohio, a few days before his untimely death.

Another biographical note has been prepared by his close friend Harry W. Trudell and appeared in *Rocks and Minerals*, **27**, 478–480 (1952).

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MEMORIAL OF MAXWELL NAYLOR SHORT

John W. Anthony, *University of Arizona, Tucson, Arizona.*

Maxwell Naylor Short died at his home in Tucson, Arizona, March 23, 1952. The shock of his death was more keenly felt because he had appeared at the time to be making a remarkable recovery from an illness that had incapacitated him during the fall months of 1951. Dr. Short's passing is felt keenly not only by his family but also by his many friends and students, and the numerous admirers of his scientific contributions to geology and mineralogy.

Dr. Short was honored in many ways for distinguished service in his

Maxwell Naylor Short

1889–1952

chosen field. He was a Fellow of the Mineralogical Society of America and was elected president in 1939. He served for many years, until the time of his death, as one of the associate editors of *The American Mineralogist*. He was a Fellow of the Geological Society of America and vice-president of that organization in 1940. He was an active member of the American Institute of Mining and Metallurgical Engineers, the Society of Economic Geologists, and of the honorary societies Tau Beta Pi and Sigma Xi. He was elected president of the Arizona chapter of Sigma Xi in 1941.

Max Short was born in Pembina, North Dakota, March 1, 1889, the son of Augustus and Eliza Jane (Naylor) Short. He was brought up on his father's farm in the Red River valley and throughout his life possessed a love of growing things. At his home in Tucson he cultivated a fine garden, specializing in roses. He spent many summers during his later years on the farm in Pembina and, among his various activities, developed a wild life preserve there.

Dr. Short completed his high school course in Los Gatos, California, and in 1911 received a degree from the University of California in mining engineering. Subsequently he followed the mining profession, working at a number of places in Arizona and Mexico. He was always fond of recounting the experiences he had during those early days. In Arizona he was employed by Phelps Dodge Corporation, the Calumet Mining Company, and the Magma Copper Company; in Mexico he was employed by El Rayo Mining and Developing Company (in Chihuahua) and the Tecolotes Mining Company. He was always proud of the fact that he had "mucked" with the best of them.

During World War I Short suffered a serious accident while serving at Camp Fremont, California, with the 319th Engineers. He lost both legs in a mine explosion. This loss handicapped him greatly but did not stop him permanently from doing a great deal of field work, especially underground. During World War II he filled in voluntarily as instructor in topographic mapping and spent entire days in the rugged terrain and heat of southern Arizona. The accident changed Short's life, but in making adjustments to new conditions he contributed greatly to geology in fields he probably otherwise would not have pursued. It is a great tribute to his perseverance and power of will that, in a profession normally followed by men of more than average physical stamina, he accomplished so much.

Following the accident of World War I, Short decided upon geology as a career and after his recovery entered Harvard University where he made an enviable academic record as a graduate student. He began a study of the opaque minerals while at Harvard and used this as a thesis for his doctorate which he received in 1923. He was appointed an instructor at Harvard during the following two years and then entered the service of the United States Geological Survey. He returned to fill the post of Lecturer at Harvard during the years 1927–28.

Dr. Short's researches both at Harvard and with the U. S. Geological Survey followed his major field of interest, the identification and genesis of the opaque ore minerals. His keenness of observation and laboratory skill are familiar to many who have used his classic work, "Microscopic Determination of the Ore Minerals." This outstanding contribution to the geological and mineralogical literature first appeared in 1931 and immediately attained a prominent place, both in the libraries of mineralogists and economic geologists, and as a classroom text. Short's original observations and methods, and his evaluation and representation of pertinent techniques from other fields stimulate the reader by their lucid presentation. A major revision of the work appeared in 1940.

Short joined the faculty of the University of Arizona in 1931, but continued on a WAE basis with the U. S. Geological Survey for many years. He became Professor of Petrography and Mineralogy at Arizona and head of the Department of Geology and Mineralogy in 1948, a position he held until forced by illness to relinquish it in 1951.

The decision to become an educator was a happy one. Not only did Dr. Short possess high scientific qualifications, but his great patience and capacity to explain with clarity difficult phases of a subject were an invaluable aid to his students. He entered into this sphere of his career as a teacher—not as a pedant. He loved his students and earnestly desired that they obtain the maximum that he was capable of giving. His classes in mineralogy and petrography were usually small, but he lectured in a formal style and with an erudite phraseology which delighed his students. This was quite at variance with his friendly informality and shyness he exhibited at other times, both at home and at the University.

One of Dr. Short's most notable characteristics was his unselfish desire to assist those with whom he came in contact. This trait explains why so many people regarded him as a close friend. He always took great pleasure each year in photographing his advanced classes and hanging the framed photographs in his laboratory. Short was also unstinting in his energies in assisting his fellow instructors, even in such menial chores as grading papers in elementary courses and tutoring undergraduates in subjects other than geology. Certain favorite expressions of enthusiasm and approval, as "textbook stuff" and "see it with the naked eye" have become a part of the geologic vocabularies of many of his students and professional associates.

Dr. Short's attitude toward academic life and his intellectual curiosity can be illustrated by the fact that he frequently audited courses of his associates. Undergraduates were sometimes startled at first to see him attending certain classes regularly and taking an active part in the discussions. He was also interested in other subjects and for many years prior to his death, he registered in the College of Fine Arts and pursued studies for the piano. He loved to perform at his home for friends.

The high integrity represented by Short's contributions to geology insure their lasting value. At the time of his death he was collaborating on a textbook on hand specimen petrology.

Dr. Short is survived by two sisters, Virginia and Eleanor (Mrs. William Norton), and a nephew, Dr. Charles A. Short, Jr.

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MEMORIAL OF CHARLES ROBINSON TOOTHAKER

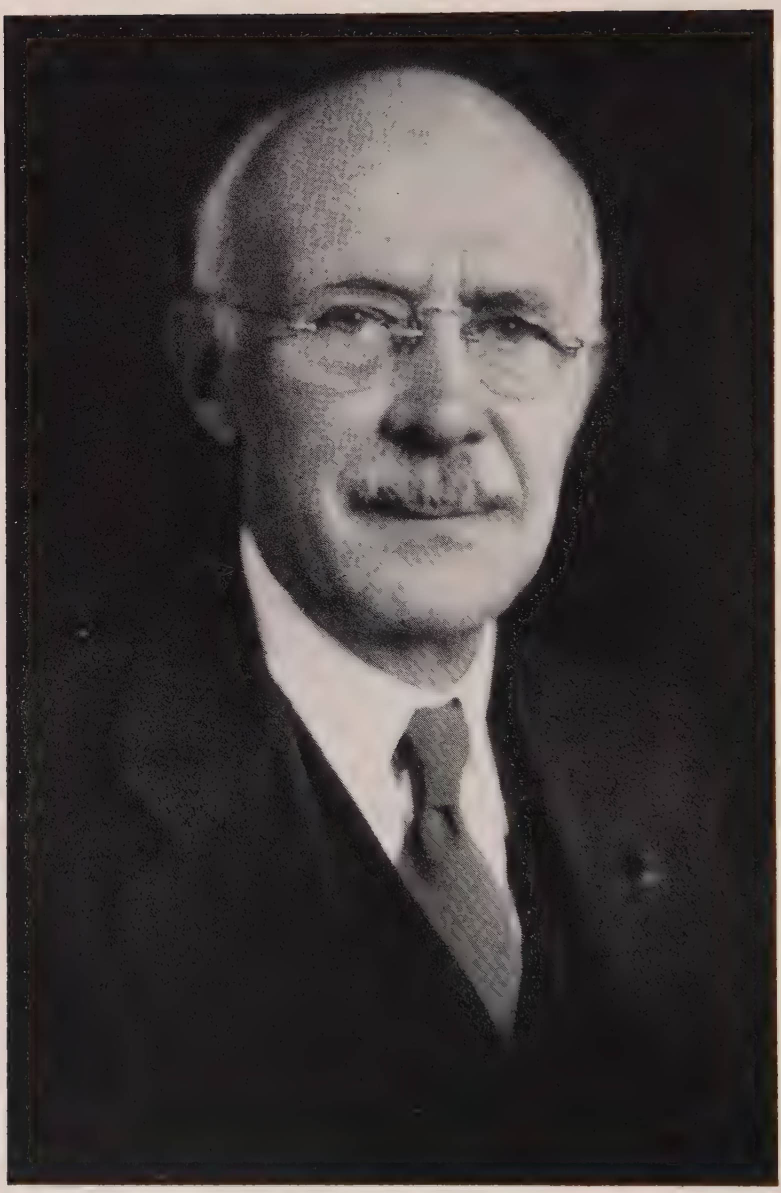
George T. Faust, *U. S. Geological Surveys, Washington, D. C.*

Charles Robinson Toothaker, Curator of the Philadelphia Commercial Museum for 54 years, died Sunday morning, May 25, 1952, in the University Hospital of Philadelphia after a month's illness. He was born in Philadelphia on May 4, 1873, of Zetta Elder and Charles Everett Toothaker, a physician. As a boy, he attended the Central Manual Training High School in Philadelphia. While in high school he took an active part in athletics, certain phases of which he continued throughout his life. After graduation from high school in 1890, he continued his education on a part-time basis. This included taking special courses in geology under Professor Edward Drinker Cope of the University of Pennsylvania and other courses of particular need in his work, for he had joined the staff of the famous Dr. A. E. Foote, mineralogist and mineral dealer of Philadelphia. Dr. Foote's company, for many years the chief supplier of mineral specimens to schools and colleges in the United States and abroad, was the forerunner of the present Foote Mineral Company, now prominent in industrial mineral research. During his work for Dr. Foote, Charles Toothaker gained a wide knowledge of topographical mineralogy and became well versed in mineral identification. In later years this knowledge was often sought by mineralogists wishing to locate original or type specimens. As a member of Foote's staff, he was assigned, as Commissioner in charge, the task of preparing the Pennsylvania Mining Exhibit of the Atlanta Exposition of 1895.

He remained with the Foote Company until 1898, when he accepted the position of Assistant Curator, in charge of exhibits, at the Commercial Museum. In 1904 his industry and ability were recognized and rewarded by his being promoted to the rank of Curator. In the years that followed, however, it was evident that his work included not only curatorial duties but administrative duties as well.

While he was Curator of the Commercial Museum, he served as advisor to industry and to the city officials of Philadelphia on problems involving the raw materials imported or shipped to Philadelphia and vicinity. In this capacity he was in a position to introduce new materials to industry, and minerals were his particular interest. He also published articles, in popular tone, which were particularly effective in bringing raw materials to the attention of possible users. Along these lines he also gained wide recognition for his lectures in commercial geography.

Charles Toothaker was particularly successful in his plan to make the Commercial Museum a vital part of the educational system. According to a press release from the Museum, "He pioneered in the movement to



Charles Robinson Toothaker

1873–1952

convert museums from inert collections of inanimate materials into living, dynamic institutions for the promotion of popular education. Mr. Toothaker projected his museum into thousands of Pennsylvania schools by means of special collections, carefully prepared for educational use. He taught the new philosophy of museum administration in widely disseminated writings. His methods and ideals, adopted in hundreds of communities, have won for museums a permanent place of usefulness in the modern world."

In connection with his work, Charles Toothaker visited and maintained cordial relations with a large number of the world's famous museums. Ever mindful of the needs of museums, he was deeply touched by the loss of the mineral collections during World War II at the University of Liege, in Belgium.

When the American mineralogists decided to come to the rescue, he volunteered to assemble their gifts and to ship them. He personally packed them with so much care that even the most delicate ones were unharmed in shipment.

His travels were also made with the welfare of the Museum in mind. In 1922 and 1923, at the invitation of the Secretary of State, he attended the Brazilian Centennial Exposition in Rio de Janeiro with his colleague Dr. William P. Wilson. He used this opportunity to travel extensively in Brazil, up the Amazon River and to important mining areas, obtaining numerous specimens for the Museum collections. In the summer of 1937 he made a collection of minerals from the cryolite deposit at Ivigtut, Greenland, which he described in *Rocks and Minerals* in 1937. His publications comprise brief notes and articles dealing chiefly with minerals and museums and one book on industrial raw materials, which has been a source of information since its initial appearance in 1905. He was a natural-born writer with a literary style that appealed especially to the average reader interested in scientific subjects.

For many years, Charles Toothaker was an active member of the Philadelphia Mineralogical Society, holding various offices. He exerted a profound influence on the junior membership of the Society and was directly or indirectly responsible for many of the members choosing to enter the geological and chemical professions, or to pursue advanced study in mineralogy. During my high school years, I was fortunate to be among those who benefited by his counsel and to be influenced in the direction of mineralogy.

Charles Toothaker was happy in the recognition which he received for his attainments in his chosen field of work. On April 24, 1930, he was elected a member of Kappa Chapter of Sigma Gamma Epsilon at Pennsylvania State College. On May 9, 1946, he was elected a Fellow of the Rochester Museum of Arts and Sciences by reason of his distinguished contributions to the field of museum education. In 1946 the Mineralogical Society of America, of which he was a member for many years, elected him a Fellow, an honor which he cherished. In 1951, he was elected a Fellow of the American Association for the Advancement of Science.

In addition to the aforementioned societies, Charles Toothaker was a member of the Pennsylvania Society, Sons of the Revolution, Schoolmen's Club, the Pan American Association, the American Association of Museums, the Museum Council of Philadelphia, and the Academy of Natural Sciences, Philadelphia.

It was typical of the man that, along with all his regular duties, he should find time for his continued participation in athletics. He was a member of the first organized basketball team in Philadelphia. At the age of 50, he was high-jumping, and at 60 and later he was still demonstrating his skill as a fencer.

From 1928 to 1952 he served as Consul in Philadelphia for Czechoslovakia, and in 1933 was made an officer of the Order of the White Lion of Czechoslovakia. He also served as Consul for Colombia.

On September 27, 1904, he married Martha Taylor McCandless, who survives him and who has been very helpful in furnishing material for the preparation of this biographical sketch. The members of the staff of the Commercial Museum, in particular Miss Marie L. Marley, have also very kindly supplied data.

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PROCEEDINGS OF THE THIRTY-THIRD ANNUAL MEETING OF THE MINERALOGICAL SOCIETY OF AMERICA AT BOSTON, MASSACHUSETTS

C. S. Hurlbut, Jr., *Secretary.*

The thirty-third annual meeting of the Society, which was held on November 13–15, 1952, at the Hotel Statler, Boston, Massachusetts, was attended by 244 Fellows and Members. Scientific sessions were held on the morning and afternoon of November 13 and 15, and on the afternoon of November 14. The Program of the Mineralogical Society carried titles and abstracts of 86 papers, 81 of which were presented orally.

The annual luncheon of the Society on November 14 was attended by 171 Fellows, Members and guests. Following the luncheon the eleventh presentation of the Roebling Medal was made to Fred. E. Wright and the second presentation of the Mineralogical Society of America Award was made to Frederick H. Stewart. The Society was addressed by the Retiring President, Michael Fleischer, on *Some Problems of Chemical Mineralogy.*

The 1952 Council of the Society met for 13 hours during November 12 and 14, 1952, and discussed 24 items of business. The 1953 Council met briefly on November 15. Much of the business of the 1952 Council was of a routine nature, but some items that are of interest to the general membership are mentioned below.

1. *Revision of the By-Laws.* It has been felt by a group of Fellows of the Society that the ballot for officers should carry two names for each office. Accordingly, the Council debated this question at length but decided against such action. However, the Council felt that, if it were made easier for nominations to be made by Fellows and Members at large, somewhat the same end would be accomplished. To achieve this purpose the Council voted to approve the following change in Article IV, Section 1 of the By-Laws and to present it to the membership at the time of the election of officers for 1954:

From

Nominations for office shall be made by the Council. The list shall be published in the Journal of the Society at least three months before the annual meeting. Any ten (10) fellows or members may forward to the secretary other nominations for any or all offices. All such nominations reaching the secretary not later than three months prior to the annual meeting shall be printed, together with the names of the nominators, as special ballots. The regular and special ballots shall then be mailed to the general membership. The results shall be announced at the annual meeting, and the officers thus elected shall enter upon duty at the adjournment of the meeting.

To

Nominations for office shall be made by the Council. The list shall be published in the Journal of the Society at least three months before the annual meeting. Any ten (10) fellows or members may forward to the secretary other nominations for any or all offices. All such nominations reaching the secretary not later than three months prior to the annual meeting shall also be printed on the ballots to be mailed to the general membership. The results shall be announced at the annual meeting, and the officers thus elected shall enter upon duty at the adjournment of the meeting.

This change would enable 10 Fellows or Members to propose nominations which would be printed on the regular ballot together with nominations made by the Council without differentiating between the two slates.

2. *Committee on Committees.* The 1951 Council appointed a committee to review the committee structure of the Society and to report to the Council at the 1952 meeting. The report of this Committee as accepted by the Council is as follows:

1. *Nominating Committee for Fellows*

It shall be the duty of this Committee to examine the qualifications of persons nominated for Fellowship in the Society and of others, qualified but not nominated, and to report each year to the Council for election. The Committee shall consist of six Fellows, appointed by the Council to serve three-year terms, two appointments to expire each year. The Chairman shall be designated by the Council.

2. *Nominating Committee for Officers*

It shall be the duty of this Committee to recommend to the Council candidates for the position of President, Vice-President, Secretary, Treasurer, Editor, and Council or Councilor (4-year term). The Committee shall consist of five Fellows to be appointed by the Council to serve one-year terms. The Chairman shall be one who has served on the Committee during the previous year.

3. *Roebling Medal Committee*

It shall be the duty of this Committee to recommend to the Council whether an award of the Roebling Medal shall be made in the following year and to recommend the recipient, if any. The Committee shall consist of three Fellows to be appointed by the Council to serve one-year terms. The Chairman shall be one who has served on the Committee during the preceding year.

4. *Mineralogical Society of America Award Committee*

It shall be the duty of this Committee to recommend to the Council whether an award shall be made during the following year and to recommend the recipient of the Award, if any. The Committee shall consist of five Fellows to be appointed by the Council to serve one-year terms. The Chairman shall be one who has served on the Committee during the preceding year.

5. *Financial Advisory Committee*

It shall be the duty of this Committee to serve as advisers to the Treasurer on matters pertaining to the investment of the Society's funds. The Committee shall consist of three Fellows or Members to be appointed by the Council to serve three-year terms. The Chairman shall be designated by the Council; he shall be one who has served on the Committee during the preceding year.

6. *Auditing Committee*

It shall be the duty of this Committee to inspect and verify the books of the Treasurer as of August 1 and to report its findings to the Council. The Committee shall consist of three Fellows or Members to be appointed by the President to serve one year. The Chairman shall be one who has served on the Committee during the preceding year.

7. *Committee on Nomenclature*

It shall be the duty of this Committee to consider problems of mineralogical nomenclature, to maintain contact with other societies or their committees, and to conduct correspondence and conferences with them aimed at reaching international agreement on such problems, and to recommend action by the Council on matters of nomenclature. The Committee shall consist of six Fellows to be appointed by the Council for three-year terms. The Chairman, designated by the Council, shall be one who has served during the preceding year. If necessary, the Council may appoint, with the advice of the Chairman, subcommittees to serve for specified times to deal with special problems.

8. *Program Committee*

It shall be the duty of the Program Committee to consider the types of program to be presented at the annual meeting of the Society, to suggest suitable subjects for Symposia at these meetings, and to make recommendations to the Council if changes in programs are deemed desirable. The Chairman shall serve as representative of the Society on the Program Committee of the Geological Society of America; the Secretary is ex officio a consultant of this Committee. The Committee shall consist of three Fellows appointed by the Council for three-year terms; the Secretary shall be a member ex officio.

9. *Editorial Board, The American Mineralogist (Board of Associate Editors)*

It shall be the duty of the Board to serve as advisers to the Editor on matters of policy pertaining to the conduct of the Society's Journal, *The American Mineralogist*. The Board shall consist of six Fellows, two to be appointed each year by the Council for a term of three years.

10. *Special Publications Editorial Board*

It shall be the duty of this Board to consider manuscripts suitable for publication in the Mineralogical Society of America Special Publications Series, to oversee their editing, and to make recommendations to the Council as to their publication. The Board shall consist of six Fellows to be appointed by the Council for three-year terms.

11. *Representative to National Research Council*

One, to serve three years, appointed by the Council.

12. *Representative to American Geological Institute*

Two, one appointed each year by the Council, to serve two years.

13. *Executive Committee*

There shall be an Executive Committee of three (President, Secretary and Treasurer) to handle matters that arise during the year and require prompt action.

14. *Committee on Committees*

There shall be a Committee on Committees composed of five Fellows. This Committee is to be appointed by the Vice-President and is to report to the incoming Council at its first meeting.

COMMITTEES OF THE MINERALOGICAL SOCIETY OF AMERICA

Roebling Medal Committee

H. H. Hess, *Chairman*
H. E. Merwin
J. D. H. Donnay

Mineralogical Society of America Award Committee

V. T. Allen, *Chairman*
T. F. W. Barth
W. F. Bradley
J. W. Gruner
R. S. Cannon, Jr.

Nominating Committee for Officers

E. W. Nuffield, *Chairman*
B. Stringham
J. R. Goldsmith
H. Winchell
J. F. Schairer

The above committees reported to the 1952 Council. It should be pointed out that these committees make recommendations, but the Council is responsible for final choice. In order to have a larger participation in the affairs of the Society, Fellows and Members are requested to send to the Secretary suggestions for candidates for Awards and for officers.

Nominating Committee for Fellows

W. Parrish and S. Zerfoss 1953
A. Buddington, *Chairman*, and H. Insley 1953–54
J. T. Lonsdale and R. B. Ferguson 1953–55

Financial Advisory Chairman

R. E. Fuller 1953
J. P. Marble 1953–54
E. P. Henderson, *Chairman*, 1953–55

Auditing Committee

H. S. Yoder, Jr., *Chairman*
(The remaining members are to be appointed later by the President.)

Nomenclature Committee

M. Fleischer and B. H. Mason 1953
C. Frondel, *Chairman*, and W. F. Bradley 1953–54
A. L. Howland and E. W. Nuffield 1953–55

Program Committee

G. T. Faust, *Chairman*, 1953
R. A. Rowland 1953–54
J. T. Goldsmith 1953–55

Special Publications Editorial Board

C. Frondel, *Chairman*, and M. Fleischer 1953
A. F. Buddington and L. S. Ramsdell 1953–54
F. A. Bannister and A. Pabst 1953–55

Board of Associate Editors
E. S. Larsen, Jr. and A. F. Rogers 1953
M. Fleischer and G. Tunell 1953–54
A. Pabst and E. F. Osborn 1953–55

Representatives
National Research Council
E. Ingerson
American Geological Institute
J. P. Marble 1953
W. T. Pecora 1953–54

REPORT OF THE SECRETARY

To the Council of the Mineralogical Society of America:

Election of Officers and Fellows

Five hundred ballots were cast in the election of officers: 148 by Fellows and 352 by Members of the Society. The officers elected to serve in 1953 are:
President: J. D. H. Donnay, The Johns Hopkins University, Baltimore, Maryland
Vice-President: H. V. Ellsworth, Canada Geological Survey, Ottawa, Canada
Secretary: C. S. Hurlbut, Jr., Harvard University, Cambridge, Massachusetts
Treasurer: Earl Ingerson, U. S. Geological Survey, Washington, D. C.
Editor: Walter F. Hunt, University of Michigan, Ann Arbor, Michigan
Councilor (1953–56) C. Osborne Hutton, Stanford University, Palo Alto, California

Dr. H. V. Ellsworth's untimely death in October 1952 left a vacancy in the office of Vice-President for 1953. To fill this vacancy the Council appointed Sterling B. Hendricks Acting Vice-President for 1953.

According to the provisions of the Constitution, the following have been elected to fellowship:

Louis Herman Ahrens, Massachusetts Institute of Technology, Cambridge, Massachusetts.
Ciro Andreatta, Universita di Bologna, Bologna, Italy.
W(illiam) M(ortimer) D(rein) Bryant, E. I. duPont de Nemours and Company, Wilmington, Delaware.
Margaret Dorothy Foster, U. S. Geological Survey, Washington, D. C.
Kenneth T(itsworth) Greene, U. S. Bureau of Reclamation, Denver, Colorado.
Anna Martta Hietanen-Makela, U. S. Geological Survey, Washington, D. C.
Harry William Trudell, Abington, Pennsylvania.
Orville Frank Tuttle, Geophysical Laboratory, Washington, D. C.
Hatten Schuyler Yoder, Jr., Geophysical Laboratory, Washington, D. C.
(Frederik) William Houlder Zachariasen, University of Chicago, Chicago, Illinois.
Peter Zodac, Peekskill, New York.

MEMBERSHIP STATISTICS

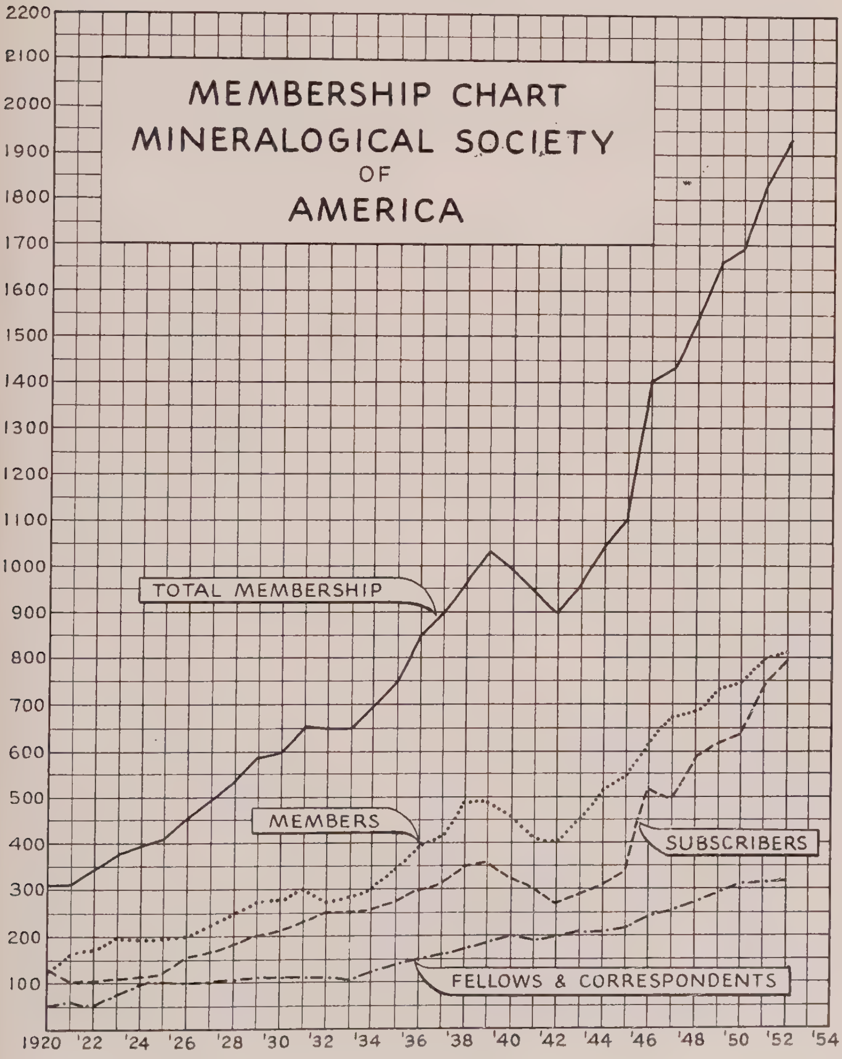
November 1, 1952

	1951	*1952*	*Gain*	*Loss*
Correspondents	5	5	0	0
Fellows	309	315	11	5
Members	797	814	143	126
Subscribers	735	799	130	66
	1846	1933	284	197

The above figures show a net gain of 6 fellows, 17 members and 64 subscribers. Considering the four groups together, there is a gain of 87, giving a total of 1933.

The Society lost through death four fellows: H. V. Ellsworth of the Canada Geological Survey, Ottawa, Canada; Samuel G. Gordon of the Atomic Energy Commission Laboratory, Los Alamos, New Mexico; Max N. Short of the University of Arizona, Tucson, Arizona; and Charles R. Toothaker of the Commercial Museum, Philadelphia, Pennsylvania.

Respectfully submitted,
C. S. Hurlbut, Jr., *Secretary*



REPORT OF THE TREASURER FOR 1952

To the Council of the Mineralogical Society of America:

Your treasurer submits herewith his report for the fiscal period beginning November 1, 1951, and ending July 31, 1952:

Receipts

Dues and subscriptions	$ 7,726.07
Sale of back numbers	731.04
Authors' charges on reprints	830.58
Interest and dividends from endowment	3,675.10
Geological Society of America aid for printing the Journal	4,914.37
Advertising	966.17
Sale of Index, volumes 21–30	54.30
Sale of Index, volumes 1–20	4.70
Securities called and sold	1,318.75
Balance from Detroit luncheon	18.50
Contribution	1.00
	$20,240.58
Cash on hand, November 1, 1951	5,601.87
	$25,842.45

Disbursements

Printing and distribution of the Journal (5 issues)	$12,784.61
Printing and distribution of reprints	1,016.68
To the Editor	375.00
Clerical assistance	426.00
Postage and express	318.33
Printing and stationery	230.88
Office supplies and equipment	156.09
1951 program and abstracts	425.00
New securities purchased	6,350.00
Commission on securities	63.66
Taxes on securities	3.23
Committee expenses	27.78
Expenses of officers to 1951 annual meeting	98.98
Safety deposit box	7.80
Refunds	32.21
Checks returned	8.00
Exchange charges on checks	1.72
Telephone and telegraph	4.16
M. S. A. Award certificates	146.00
	$22,476.08
Cash balance, July 31, 1952	3,366.37
	$25,842.45

The endowment funds of the Society as of July 31, 1952, consist of the following securities:

Bonds

6M Atlantic Coast Line, 4½	$5,257.50	
5M New York Central, 5	4,300.00	
5M Southern Railway	5,743.75	
		$15,301.25

Preferred Stocks

200 shares, Southern California Edison, 4.88	$5,250.00	
100 shares, Union Pacific, 4	4,570.25	
60 shares, Jones and Laughlin, A, 5	4,987.50	
55 shares, United States Steel, 7	6,946.20	
50 shares, Virginia Electric & Power Co., 5	5,942.50	
24 shares, Public Service Electric & Gas Co.	728.40	
10 shares, Consolidated Edison, 5	1,066.64	
		$29,491.49

Common Stocks

288 shares, Potomac Electric Power Co.	$3,871.00	
200 shares, Greyhound Corp.	2,300.00	
150 shares, Spencer Kellog	3,775.00	
100 shares, Columbus & Southern Ohio Electric	2,087.50	
80 shares, Kroger Co.	1,990.00	
60 shares, United Fruit Co.	3,067.50	
56 shares, Standard Oil of New Jersey	1,444.84	
50 shares, Chesapeake & Ohio Railway	2,368.75	
50 shares, Phelps Dodge	1,975.00	
47 shares, American Telephone & Telegraph	6,427.32	
A. T. & T. Convertible Debentures	600.00	
40 shares, Plymouth Cordage	2,050.00	
30 shares, U. S. Playing Card	2,411.25	
		$34,368.16
		$79,160.00

Respectfully submitted,
EARL INGERSON, *Treasurer*

DANA FUND

Disbursements are made to needy mineralogists in war areas and to families of deceased mineralogists in war areas.

Receipts

Available balance, November 1, 1951	$170.96
Interest	2.45
	$173.41

DISBURSEMENTS

Disbursed	$ 25.00
Available balance, July 31, 1952	148.41
	$173.41

Respectfully submitted,
EARL INGERSON, *Treasurer*

REPORT OF THE AUDITING COMMITTEE

To the President of the Mineralogical Society of America:

The Auditing Committee has examined and verified the accounts of the Treasurer of the Mineralogical Society of America for the fiscal period beginning November 1, 1951, and ending July 31, 1952. The securities listed in the Treasurer's report, with all future coupons on the coupon bonds attached, are in the safety deposit box at the West End Branch of the Washington Loan and Trust Company in Washington, D. C.

Respectfully submitted,
HOWARD W. JAFFE
HATTEN S. YODER, JR.
MARIE L. LINDBERG, *Chairman*

REPORT OF THE EDITOR FOR 1952

To the Council of the Mineralogical Society of America:

In presenting the annual editorial report for 1952 at this time, the same general policy will be followed as that used in recent years. With five issues published and distributed and the November-December number in press, sufficient data are now available to give a fairly accurate general survey of the year's activity. When the editor's report appears in print in the March-April issue certain details incomplete or not available at present will be incorporated so that the final printed report in April can be compared with those given for previous years.

If the success of a year's activity can be judged by the number of articles published and by the total pagination, then the accomplishments for 1952 can be considered well above the average. The current volume will run close to 1090 pages, an increase of approximately 150 pages, or 16%, above the figures achieved last year.

Aside from the five regular issues, one special number, authorized by the Council, and sponsored by the Walker Mineralogical Club, appeared as the May-June number. This particular issue continued the series of Contributions to Canadian Mineralogy under the able editorship of L. G. Berry of Kingston, Ontario. In this number the guest editor assembled 17 interesting papers on varied mineralogical subjects and assumed full responsibility in seeing these contributions through the press. This policy, followed for the past four years, of devoting one issue each year exclusively to Canadian Mineralogy and Canadian authors seems to have won general approval, and by Council action will be continued in 1953.

Attention should also be called to a colored plate (2 figures) appearing in the November-December number on the "Effect of heavy charged particle and fast neutron irradiation on diamonds." One of the authors was so anxious to have this effect reproduced in color that he personally will assume the entire cost of these colored reproductions.

The current volume will run, as indicated above, about 1090 pages and contain 77 leading articles. Twenty-three additional short papers, appearing under Notes and News, give an overall total of 100 published manuscripts for the year. These contributions were re-

TABLE 1. DISTRIBUTION OF SUBJECT MATTER IN VOLUME 37

Subjects	Articles	Pages	Per Cent of Total
Leading articles*			
Descriptive mineralogy	14		
Geochemistry	23		
Structural crystallography	20		
Petrography	5		
Optical crystallography	4		
Mineralography	2		
Geometrical crystallography	1		
Memorials	4		
Miscellaneous	4		
	77	892	83.0
Shorter articles	23	$55\frac{1}{2}$	17.0
Notes and news	35	$13\frac{1}{2}$	
Proceedings of Societies	4	65	
Abstracts of new mineral names	39	11	
Book reviews	34	37	
Total entries	212	1074	100.0
Illustrations	291		
Index, Title page, Table of contents		20	
Grand total		1094	

* Leading articles averaged 11.6 printed pages each.

ceived from 127 authors associated with 57 different Universities, research bureaus, and technical laboratories. Thirty-four critical book reviews, thirty-nine abstracts of new mineral names, and four new minerals described in detail for the first time—bystromite, aurostibite, robinsonite, and hurlbutite—are a few of the other items of general mineralogical interest.

It should be a matter of considerable pride to the Society to note that the service rendered by the Journal is world wide in its scope. As an indication of this fact, during the past year 31 contributions (out of 100) were received from contributors residing outside of the States; 17 from Canada; three each from Australia, South Africa, and Japan; and one each from England, Formosa, New Zealand, India, and Venezuela.

While it is impossible at this early date to state with any degree of extreme accuracy the printing demands that will be made on the Journal in 1953, it would seem from present indications that no depression is in sight, and that the volume for 1953 conservatively speaking might easily reach the 1200 page level. This estimate is based on the fact that as of November 1 the files contain 31 articles of various lengths aggregating about 700 typed pages. This estimate does not include one or more special numbers that are contemplated, whose lengths cannot be estimated at this time.

This report would not be complete without an expression of appreciation for the liberal and substantial financial support received from the Geological Society of America toward defraying a sizeable portion of the printing costs of the Journal. For this assistance the Society is indeed very grateful for without such aid expansion would be impossible and services of necessity would have to be curtailed.

The accompanying Table 1 summarizes in more detail the distribution of subject matter in volume 37.

Respectfully submitted
Walter F. Hunt, *Editor*

Bibliography on Uranium in Arizona, Nevada, and New Mexico

The Geological Society of America published the following article in the February, 1953, issue of its Bulletin: "Bibliography and Index of Published Literature on Uranium and Thorium and Radioactive Occurrences in the United States. Part 1: Arizona, Nevada, and New Mexico," by Margaret Cooper of the Division of Raw Materials, U. S. Atomic Energy Commission. Since this 38-page bibliography may prove helpful to both geologists and laymen interested in uranium prospecting, the Society has prepared reprints for public sale at 25 cents per copy. Remittance must accompany orders, which should be sent to:

The Geological Society of America
419 West 117 Street, New York 27, New York

Opening for Assistant Professor of Mineralogy

The Ohio State University has a vacancy in the Department of Mineralogy for an Assistant Professor. In order to be considered, applications should be submitted prior to June 1, 1953, to Professor Duncan McConnell, *Chairman.*

Dr. Edward Steidle, Dean of the School of Mineral Industries at the Pennsylvania State College since 1928, will retire with emeritus rank on June 30. He will be succeeded by Dr. Elburt F. Osborn, associate dean of the School. Dr. O. F. Tuttle has been named professor of geochemistry and Head of the Department of Earth Sciences, effective July 1, 1953. Dr. John A. Hipple, Chief of the Atomic Section, National Bureau of Standards, Washington, D. C., has been named Director of the Mineral Industries Experiment Station.

Erratum

Am. Mineral., **37** (1952) p. 598 last line: read *Fig. 2* instead of *Fig. 1.*

ABSTRACTS OF PAPERS PRESENTED AT THE THIRTY-THIRD ANNUAL MEETING OF THE MINERALOGICAL SOCIETY OF AMERICA, NOV. 13–15, 1952

COMPARISON OF THE CHEMICAL ANALYSES OF SEDIMENTARY AND METAMORPHIC ROCKS

ARDEN L. ALBEE
Montpelier, Vermont

The composition of a metamorphic rock is commonly compared with that of the supposed parent rock to determine if chemical changes have taken place during metamorphism. Chemical analyses of metamorphic rocks supposedly derived from sedimentary rocks are frequently compared with composite analyses of sedimentary rocks (commonly Clarke's), and certain compositional features are compared with those that have come to be accepted as generally characteristic of sediments. E. S. Bastin has cited the dominance of MgO over CaO, the dominance of K_2O over Na_2O, an excess of Al_2O_3 over the amount necessary for feldspars, and a high SiO_2 content as characteristic of pelitic sediments. These generalizations concerning pelitic sediments have come to be regarded by many geologists as applicable to sediments in general, and some recent textbooks have so treated them.

Complete analyses of individual argillaceous and arenaceous rocks are presented graphically to show the range of variation from the composite analyses. The graphs also show the departure of various rock types from the compositional features mentioned above.

Comparison of an analysis of a metamorphic rock with composite or averaged analyses of sedimentary rocks is shown to be an unreliable basis for hypothesizing chemical change during metamorphism. Comparison should instead be made with individual analyses of sedimentary rock, taking into account the geologic setting of the rocks to be compared.

Some of the important factors involved in the comparison of a sedimentary rock and its metamorphic equivalent are discussed.

MINERAL AGE MEASUREMENT: MASS SPECTROMETRIC DETERMINATIONS OF Rb^{87} AND Sr^{87} IN LEPIDOLITES

L. T. ALDRICH, J. B. DOAK AND G. L. DAVIS
Department of Terrestrial Magnetism and Geophysical Laboratory, Carnegie Institution, Washington, D. C.

Isotopic determinations of Rb^{87} and Sr^{87} in lepidolites have been made using stable isotope dilution and mass spectrometric techniques. Chemical procedures outlined in the paper by Davis and Aldrich yielded sufficient Sr for this analysis if 2 grams of the lepidolite were available and it was as old as 10^7 years. Rb from 100 mg of lepidolite could be separated from the K with sufficient purity for isotopic analysis with a single recrystallization of the perchlorates. In the analysis of each mineral sample 50 micrograms of Sr containing 64 per cent Sr^{84} and 180 micrograms of Rb containing 89 per cent Rb^{87} were added when the sample was first put in solution. The ratio of Sr^{87}/Rb^{87} can be determined with a precision of 3 per cent under the least favorable conditions.

Ages have been determined in this manner for five lepidolites. They range from $158 \pm 10 \times 10^6$ years for a Pala, California, sample to $3300 \pm 300 \times 10^6$ years for a specimen from Southern Rhodesia. The absolute ages depend on the decay constant of Rb^{87}, the latest published value for which is $1.09 \pm .03 \times 10^{-11}$ year^{-1}.

X-RAY EVIDENCE OF THE NATURE OF CARBONATE-APATITE

Z. S. ALTSCHULER, E. A. CISNEY, AND I. H. BARLOW
U. S. Geological Survey, Washington, D. C.

The existence of carbonate-fluorapatite as a distinct variety of apatite rather than a mixture of a carbonate phase and fluorapatite has been tested by the following x-ray experiments: (1) Several samples each of fluorapatite and hydroxyl-apatite, of approximately end-member composition, and of carbonate-fluorapatite (having 2 to 4 per cent CO_2) were x-rayed by powder-film methods under identical conditions of instrumentation and sample preparation. The resultant film patterns are distinctive and can be correlated visually with the above chemical types by reference to the (231) and (004) lines. (2) Mixtures of these several varieties were x-rayed and line duplication resulted. (3) Mixtures of two samples of the same variety were x-rayed and line duplication did not occur. (4) Unit-cell constants, computed from x-ray spectrometer patterns with an accuracy and precision of ±0.002 Å, offer proof that carbonate-fluorapatite is sufficiently different from fluorapatite and hydroxyl-apatite to be considered another variety of apatite. The cell constants follow:

	a_0 (A)	c_0 (A)	c/a
Hydroxyl-apatite, Holly Springs, Ga., U.S.A.	9.413	6.875	0.730
Fluorapatite, Durango, Mexico	9.386	6.878	0.733
Carbonate-fluorapatite, Staffel, Germany	9.344	6.881	0.736

The apatite in the marine phosphorites of Khouribga, French Morocco, of Florida and Idaho, U.S.A., and many other areas is carbonate-fluorapatite.

MAGMATIC AND GRANITIZED ROCKS IN THE YELLOWJACKET DISTRICT, LEMHI COUNTY, IDAHO

ALFRED L. ANDERSON
Cornell University, Ithaca, New York

The pre-Cambrian Belt strata in the Yellowjacket district, Idaho, have been invaded by three unrelated groups of intrusive rocks, one of which has induced local granitization. The earliest group (pre-Cretaceous) is represented by small bodies of considerably altered gabbro; the second (late Cretaceous or early Tertiary), by dikes and stocks of diabase, olivine gabbro, augite-hornblende-biotite diorite, hornblende-biotite diorite, pegmatite, and lamprophyre, and marginal masses of biotite diorite and syenite about the body of hornblende-biotite diorite and a small dike of pulaskite remote from the diorite; and the third (Miocene), by dikes and less regular bodies of granophyre, granite porphyry, and vitrophyre.

Except for the biotite diorite, syenite, and pulaskite, the rocks have all the textural and structural characteristics of more or less quickly cooled hypabyssal intrusives, complete with various reaction phenomena and in part with micrographic and microspherulitic intergrowths. The exceptions are products of alkali-rich granitizing fluids from the hornblende-biotite diorite magma, with the biotite diorite and syenite formed by replacement of the bordering Belt rocks and the pulaskite by consolidation of the fluids along more distant channelways. Metallization accompanied the second group of intrusives, with ore deposition just ahead of the injection of lamprophyres, which brought the magmatic cycle to a close. The two earlier groups of intrusives apparently had their sources in the basic subcrust; the younger in magma that had reached a higher level in the crust. None of the intrusives is related to the deeply underlying Idaho batholith.

PITCHBLENDE DEPOSITS ON QUARTZ HILL, CENTRAL CITY DISTRICT, GILPIN COUNTY, COLORADO

F. C. ARMSTRONG
Denver, Colorado

The Central City district, in the northern half of the Colorado Front Range mineral belt, has been the principal domestic source of pitchblende. A recorded 110,757 pounds of U_3O_8 has come from Quartz Hill—more than 98 per cent from the Wood and Kirk mines.

Quartz Hill is underlain by a folded sequence of pre-Cambrian biotite granite gneiss, amphibolite, quartz-biotite schist, granulite, and skarn intruded by pegmatite, metagabbro, and Upper Cretaceous bostonite. These rocks are cut by steeply dipping, east-trending faults that contain metalliferous veins. The Central City district is zoned with pyritic gold ores in the center and lead-zinc-silver ores on the periphery. Quartz Hill is on the west side of the district, and is in a transition zone where the mineralogy of the veins changes from quartz-pyrite-gold on the east to quartz-pyrite-sphalerite-chalcopyrite-galena-gold-silver on the west. Pitchblende occurs with chalcopyrite in this transition zone. Gold ore bodies are grossly controlled by the intersection of faults with brittle granitic layers, and pitchblende ore bodies probably have a similar control. Mineralogically and structurally the veins are of xenothermal type.

Ore produced from the Wood and Kirk mines had an estimated average grade of 0.2 per cent U_3O_8. It is probable that unmined parts of the Wood and Kirk veins within the transition zone contain ore of similar grade.

RADON LEAKAGE FROM RADIOACTIVE MINERALS

GEORGE L. BATE, BRUNO J. GILETTI, AND J. LAURENCE KULP
Lamont Geological Observatory, Palisades, New York

The rate of radon leakage from selected radioactive minerals and ores has been measured as a function of particle size and temperature. The results suggest that this factor is a significant correction for certain types of minerals used in uranium-lead age measurement. For disseminated types of ores such as occur at the Sunshine Mine, Idaho, the correction may be as large as 20% of the Pb^{206} content. Certain massive ores or clean crystals show leakage rates on the order of 1% or less. The leakage rate is not largely affected by temperature until several hundred degrees are reached. This suggests that an integrated radon leakage for the entire history of the mineral can be calculated with adequate precision for the age measurement. The method by which such a calculation is made and the error involved in the final age value by this procedure will be discussed for a particular case.

PLEONASTE FROM CABALLO MOUNTAINS, NEW MEXICO

CARL W. BECK AND CHARLES H. MAXWELL
University of New Mexico, Albuquerque, New Mexico

Pleonaste crystals occur as large phenocrysts in several nepheline olivine basalt flows at the north end of the Caballo Mountains, New Mexico. These Quaternary (?) flows overlie Cretaceous and Permian strata and are separated from them by an angular unconformity.

The pleonastic spinels from this area range in size from microscopic up to 15 mm. Most of the pleonaste crystals are anhedrons or malformed octahedrons. Good octahedrons, or octahedrons truncated by dodecahedrons, are not uncommon, and one cubic truncation has been noted. The octahedron faces are usually pitted. The color is dark green to brownish green, nearly opaque; fracture, conchoidal; luster, dull on external surfaces, vitreous to splendent on fractures; hardness, 8; parting, cubic. The measured specific gravity ranges from 3.83 to 3.88; calculated specific gravity is 3.91. The space group is $Fd3m$, $a_0=8.12$ Å. The index of refraction ranges from 1.788 to 1.789. The empirical formula, as determined from a chemical analysis of 4.025 gm, is: $(Fe_{.35}Mg_{.65})O \cdot (Fe_{.02}Al_{.98})_2O_3$.

From petrographic studies it is concluded that the original mafic magma, before eruption, cooled and remained uniform long enough to allow large crystals of pleonaste and olivine to form, then was ejected and cooled too quickly for the liquid to resorb these early-formed crystals.

STUDIES OF METAMICT MINERALS (II): RE-EXAMINATION OF FERGUSONITE

JOSEPH BERMAN
U. S. Bureau of Mines, College Park, Maryland

With one possible exception all descriptions of fergusonite found by the author indicate that the mineral occurs only in the metamict state.

Fergusonite was investigated by T. F. W. Barth in 1926. He concluded that the mineral crystallizes at about 400° C. but the crystallized particles are too small for good *x*-ray diffraction unless heated to a temperature between 800° C. and 900° C.

My investigation shows that fergusonite is dimorphic. At approximately 450° C. metamict fergusonite crystallizes and can be indexed with orthorhombic axes. At higher temperatures (ranging from about 850° C. to 1050° C., depending on the kind of sample and the length of heating time) it crystallizes in a form identical with that of synthetic "fergusonite." Barth has indexed this synthetic "fergusonite" with a tetragonal lattice.

It is believed that the structure of the low-temperature dimorph of fergusonite is equivalent to the structure of the original crystalline material.

NEW DATA ON LEAD SULPHARSENIDES FROM BINNENTAL, SWITZERLAND

L. G. BERRY
Queen's University, Kingston, Ontario

Weissenberg and precession camera studies on baumhauerite, dufrenoysite and rathite, lead to the following new structural data:

Baumhauerite is triclinic (pseudomonoclinic) with $a=22.74$, $b=8.33$, $c=7.89$ A, $\alpha=90°$, $\beta=97°25'$, $\gamma=90°$. The unit cell contains $4[3PbS \cdot 2As_2S_3]$.

Dufrenoysite is monoclinic, probable space group $P2_1/m$; $a=8.41$, $b=25.85$, $c=7.88$ A, $\beta=90°30'$. The unit cell contains $8[2PbS \cdot As_3S_3]$ with specific gravity 5.53 (measured), 5.61 (calculated). The axial ratio of the unit cell, $a:b:c=0.326:1:0.306$ agrees closely with the morphological ratio given by Palache in Dana (1944).

Rathite-II is monoclinic, probable space group $P2_1/m$; $a=8.32$, $b=70.9$, $c=7.91$ A, $\beta=90°$. The unit cell probably contains $2[19PbS \cdot 13As_2S_3]$ with specific gravity 5.42 (measured, Jackson), 5.52 (calculated). The unit cell dimensions do not agree with previous structural or morphological data on rathite, although they were obtained on a fragment of the original material analyzed by Jackson (analysis 3, Dana, 1944) which was kindly supplied by Dr. Bannister from the British Museum (Natural History).

RELATIONSHIP BETWEEN LIGHT ABSORPTION AND COMPOSITION IN A SOLID SOLUTION SERIES

F. DONALD BLOSS
University of Chicago, Chicago, Illinois

Using a spectrophotometer improvised from a monochromator, polarizing microscope, and a photomultiplier tube, light absorbance measurements were made for the γ' direction on a (110) face for crystals of varying nickel content in the solid solution series between $Ni(NH_4)_2(SO_4)_2 \cdot 6H_2O$ and $Mg(NH_4)_2(SO_4)_2 \cdot 6H_2O$. Measured absorbance values were found to confirm the Beer-Lambert law within the limits of measurement. On the basis of this law, compositions predicted from absorbance per centimeter values of different crystals were, on the average, accurate to within 2.5 mol per cent (of the compositions predicted from density values).

Calculations to obviate the effects of opaque inclusions and internal cleavages in the measured crystals when predicting composition from absorbance measurements are discussed.

CHARACTER AND ORIGIN OF THE ROAN AND CAROLINA GNEISSES IN THE SPRUCE PINE PEGMATITE DISTRICT, NORTH CAROLINA

DONALD A. BROBST
Spruce Pine, North Carolina

A belt of upper-middle-grade metamorphic rocks that have long been divided into Roan gneiss (essentially hornblende gneiss and schist) and the Carolina gneiss (essentially mica gneiss and schist) constitute a part of the Spruce Pine pegmatite district in the Blue Ridge of western North Carolina. Detailed field mapping of 250 square miles in that district indicates that these two rocks are interlayered and the individual layers range from an inch or less to several tens of feet in thickness. The hornblende and mica rocks merge laterally and vertically.

All the metamorphic rocks have been plastically deformed into complexly folded structures. Foliation is parallel to the layers. No crosscutting relationships have been found between the Roan and Carolina gneisses.

The mica rocks include mainly biotite and muscovite gneiss and schist. The feldspar is generally oligoclase; orthoclase and microcline are rare. The schists typically have more mica, with muscovite dominating over biotite; they also have more garnet and are more distorted than the gneisses. Typical accessory minerals in the Carolina gneiss are garnet, staurolite, kyanite, epidote, allanite, apatite, zircon, and magnetite-ilmenite.

The hornblende rocks are chiefly gneiss and schist which, in addition to the hornblende, contain quartz, andesine, epidote, and magnetite-ilmenite. Associated rocks include layers rich in actinolite, epidote, zoisite, garnet, diopside, and rarely dolomite.

Field evidence and laboratory studies suggest that the Roan-Carolina complex of this area is of sedimentary origin. The Roan gneiss may have been derived from ferruginous carbonated shales and dolomitic limestones. The Carolina gneiss may have been derived from clayey to sandy shales and sandstones.

SOME ASPECTS OF THE GEOCHEMISTRY OF POTASSIUM, RUBIDIUM, CESIUM, AND THALLIUM IN SEDIMENTS

FRANK C. CANNEY
Massachusetts Institute of Technology, Cambridge, Massachusetts

Determinations of the rubidium, cesium, and thallium contents of 323 samples from 10 sedimentary geological formations have been made by spectrochemical methods. Potassium, present as a structure constituent in a known concentration, was used as a variable internal standard. The precision of the spectrochemical methods employed, expressed as standard deviations, was $\pm 4.5\%$ for rubidium, $\pm 8.2\%$ for cesium, and $\pm 8.8\%$ for thallium. Average abundance values for these elements in argillaceous sediments are: Rb_2O—0.031%, Cs_2O—0.00081%, and Tl_2O—0.000038%. Thallium is usually enriched in highly organic sediments.

The average % K_2O/% Rb_2O, % K_2O/% Cs_2O, % K_2O/% Tl_2O, and % Rb_2O/% Tl_2O abundance ratios in the various classes of sedimentary materials will be given and their significance discussed. A comparison of these values with the average values of corresponding ratios in igneous rocks shows that the order of relative enrichment of these elements in sediments is normally: $Cs > Rb > K > Tl$.

The conclusion is reached that the order of relative enrichment found is in excellent agreement with the theory that the adsorption affinity of a cation from aqueous solutions is [believed to be] proportional to the value of the positive electrostatic potential existing at the surface of the hydrated ion.

ORIGIN OF GRANITIC GNEISS DOMES OF WESTERN NEW HAMPSHIRE

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At least twelve domical structures lying within a narrow belt in western New Hampshire have been mapped and described by different investigators. Each dome consists of a central elliptical core of weakly foliated gneiss, mantled by concentric belts of metamorphosed sedimentary and volcanic rocks. The foliation of the gneissic core is most pronounced near the borders where it parallels the contact and the bedding-plane schistosity of the adjacent metamorphic rocks.

Various magmatic origins have been proposed for these Oliverian domes including the laccolithic, phacolithic, and sill theories. Virtually no consideration, however, has been given to other possible origins for these plutonic features. That metamorphic and metasomatic phenomena may have played dominant roles in the formation of these gneissic domes seems highly probable and warrants special attention. This possibility is discussed and evidence for replacement in the dome rocks is given. The time and space relation of the Oliverian domes to the general sequence of tectonic and petrologic events in central New England is considered to have a most significant bearing on the origin of the domes themselves.

It is concluded, tentatively at least, that, whereas many observations may be interpreted as indicating a magmatic origin for the gneissic cores of the Oliverian domes, our present knowledge of these plutons may be explained more logically by the metamorphic-metasomatic theory.

ISOTOPIC CONSTITUTION OF RADIOGENIC LEADS AND THE MEASUREMENT OF GEOLOGICAL TIME

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A review of all the age determinations carried out by the authors is described. Isotopic abundances and ages are listed for over 70 radiogenic leads extracted from uranium-bearing samples from different localities. Chemical analyses are given for some of the samples. Ages have been calculated from ratios of radiogenic Pb^{207}/Pb^{206}, and from the ratio of Pb^{206}/U^{238} for chemically analyzed samples.

The reliability of the chemical and isotopic ages is discussed. It is pointed out that discrepancies between the ages by the two methods may be explained by the loss of uranium or lead in recent time which may occur by partial dissolution or alteration of the sample. Evidence is presented showing that the loss of radon is small or does not occur at all, and that therefore the isotopic age values are the most reliable as was pointed out by Nier.

From age determinations combined with mineralogical evidence the behavior of hydrothermal pitchblende deposits is discussed. Evidence is given of the dissolution and redeposition of such deposits. Ages of uraninites, and of a few pitchblendes are considered useful in dating orogenies, but many pitchblendes give younger dates for that reason.

The preparation and analysis of samples is described. Quantities of lead as small as 5 mg. have proved sufficient for the preparation of adequate samples of lead tetramethyl for analysis in a mass spectrometer.

ISOTOPIC CONSTITUTION OF ORE LEADS AND THEIR RELATION TO GEOLOGICAL TIME

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Since the publication of Nier's determinations of the isotopic constitution of leads from lead ores of different ages, his data have been the basis for several estimates of the age of the earth's crust and of the maximum time of formation of the elements. All these investigators have pointed out the need for more isotopic analyses of naturally occurring leads of known age. The authors report the isotopic constitution of 14 samples of lead ores, 12 of which were collected from Precambrian areas. A number of these leads have been dated, by isotopic analyses of associated radiogenic leads by the authors, by isotopic analyses of radiogenic leads by Nier, and by Ahren's ages from the Rb-Sr method. These new values for the isotopic constitution of ordinary leads have been combined with those of Nier and used to calculate the time of solidification of the earth's crust, for which the authors report a value of 3.5 ± 0.2 billion years. This value agrees with estimates reported by Holmes, and by Bullard and Stanley within the limits assigned. A new estimate of 5.50 ± 0.2 billion years is given as a maximum limit for the time of formation of the elements.

The authors discuss the dating of ore leads from their isotopic constitution, and point out that while no general criterion may be used for all samples, it is usually possible to approximately date the separation of the ore from the parent magma.

MINERAL AGE MEASUREMENT: CHEMICAL SEPARATIONS OF Rb AND Sr BY MEANS OF ION-EXCHANGE COLUMNS

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The chemical separation of trace amounts of any element, in pure form, from minerals is a formidable problem by conventional analytical procedures. For mineral age measurements by mass-spectroscopic analyses it is particularly important to obtain strontium, rubidium, and calcium from micas, feldspars and accessory minerals from granites. The strontium must be completely separated from rubidium and usually partially separated from calcium.

Ion-exchange resin columns have been adapted to this separation and procedures developed in which, first, the alkali metals and alkaline earths are completely separated from the aluminum, iron, zirconium and other elements in the mineral, with complete recovery. Next, the alkalis are removed from the column, separated partially from each other and completely from the calcium and strontium. Finally the alkaline earths are removed with at least a partial separation.

TEXTURAL RELATIONSHIPS IN THE SALINA SALT OF MICHIGAN

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Petrographic examination of the salt bed at the base of the Salina was directed toward the study of the textural relationship between halite, anhydrite and dolomite and related pyrite, sylvite, celestite, quartz and carbonaceous material. The core examined was taken from Newago County, Michigan, the base of the salt being intersected at 5147 feet. Study of these relationships was facilitated by the frequent interbedding of dolomite and anhydrite with salt in the lower portion of the section.

Some of the particularly significant relationships are (1) variations in grain size and shape of dolomite and anhydrite, both tending toward increased size and euhedral shape in the presence of halite, (2) reversal of the normal sequence of deposition with changes upward progressing from salt to anhydrite to dolomite to salt, (3) concentration of pyrite and carbonaceous material along the contact between halite and the overlying anhydrite, (4) brecciation of anhydrite layers, the fragments being contained in a matrix of dolomite, and (5) development of a chevron structure in halite due to the orientation and concentration of liquid inclusions in the form of negative crystals.

DIRECTIONAL GRINDING HARDNESS IN DIAMOND

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In order to measure the relative variation of grinding hardness in single diamond crystals, two mechanisms have been constructed. One, which might be called a hardness comparator, permits the simultaneous cutting of two crystals. One crystal is oriented in an arbitrarily chosen unit hardness direction. The cutting rate of the second crystal is compared with the cutting rate of the reference crystal.

To facilitate the measurements a second apparatus has been designed to prepare spherical surfaces of predetermined radii on the diamonds. This apparatus was used in connection with hardness variation measurements on the cube face. The remainder of the planes investigated were studied on pyramidally faceted stones, because of the very slow operation of the sphere grinder.

Hardness curves were plotted, hardness against grinding azimuth, for the cube, dodecahedron, and octahedron, and for intermediate planes in the zones containing these three planes.

Hexoctahedral symmetry was assumed since no evidence of lower symmetry has been obtained from hardness measurements.

Errors inherent in the method are discussed.

COMPOSITIONAL CONTROL OF THE DIMENSIONAL PROPERTIES OF CRYSTALS

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The compositional control of crystal dimensions is illustrated by the Ca-containing amphiboles. The dimensions as expressed by the ratio of $\frac{\text{length in the } b \text{ direction}}{\text{length in the } c \text{ direction}}$ varies in the crystals measured from 0.09 for tremolite to 0.78 for Fe-rich hastingsite. These ratios are interpreted as being the crystallographic (or growth) expressions of the changes in bond forces as the composition changes from Al, Fe-poor tremolites to Al, Fe-rich hastingsites.

Al which substitutes for Si weakens the Si-O bonds in the *c* direction and strengthens the Metal-O bonds in the directions of the *a* and *b* axes. Fe which substitutes for Mg, and Al which substitutes for Fe and Mg further strengthen the bonds in the *a* and *b* directions and thus influence the development of the crystal in these directions.

The presence and development of the 010 face seems to be correlated with the substitution of Fe and/or Al for Mg.

Compositional control of crystal dimensions of other amphiboles and of pyroxenes is discussed. Implications of the dimensional ratios on grain size and the tendency to recrystallize are indicated.

SEISMIC REFRACTION MEASUREMENTS IN THE GULF OF MAINE

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Seismic refraction measurements were made in the Gulf of Maine in 1948 and 1952 as a continuation of the program of geophysical exploration of the continental margins. Sections from Portland, Maine, to the northwest edge of Georges Banks, from Matinicus Rock, Maine, to Cultivator Shoal, and from Cape Ann, Mass., to Yarmouth, Nova Scotia, were observed.

Sediments of 5,025 to 6,800 ft/sec varied erratically in thickness from 0 to 2,000 ft. To the north and east of Ammon Rock consolidated sediments of 12,000 to 13,000 ft/sec were detected with thicknesses up to 1,600 ft. These are tentatively identified as Triassic and possibly represent a continuation of the Fundy Basin. Basement rocks with velocities of 16,500 to 18,000 ft/sec were observed to pinch out rapidly away from shore on all sections. Sub-basement rocks with velocities of 19,000 to 21,000 ft/sec were found to underlie the entire Gulf of Maine.

PURIFICATION AND PROPERTIES OF MONTMORILLONITE

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Descriptive mineralogical data have been obtained on ten montmorillonites from bentonite deposits located in the United States and Canada using thermal, *x*-ray, electron, and chemical methods.

The montmorillonites were hand-picked and subsequently mechanically purified by settling and supercentrifugation. After purification cristobalite was detected in four of the ten samples. Lepidocroite and hydrous mica were present in one sample. The differential thermal analysis curves show endothermic peaks at (140–160° C.), (650–715° C.), and (820–885° C.) followed by an exothermic reaction above 870° C. *X*-ray powder photographs show basal spacings up to the tenth order yielding $d_{(001)} = 17.9$ to 18.0 A for glycerol treated specimens. The electron micrographs reveal small, irregular thin plates. One sample showed some of the plates curled into tubes. The base-exchange capacity varies from 66 to 128 milliequivalents per 100 grams of sample depending on impurities and chemical composition of the montmorollonite, but independent of particle size. Formulae calculated from the chemical analyses show a fairly wide range of isomorphism. The peak temperatures of the thermal reactions, the "*a*" and "*b*" dimensions of the unit cell and base-exchange capacities are correlated with the substitution of magnesium and iron for aluminum in the octahedral layer and aluminum for silicon in the tetrahedral layer as determined from chemical analyses.

THERMAL, DEHYDRATION, AND *X*-RAY STUDIES ON MONTMORILLONITE

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The thermal reactions, water loss and changes in crystal structure have been investigated between 500° C. and 950° C. for five montmorillonite samples with varying chemical composition.

The water loss during the first high temperature endotherm accounts for the water proposed by the Hogmann-Endell-Wilm structure. This reaction involves changes in the spacings of the *x*-ray diffraction lines which probably reflect symmetry differences. There is some additional water loss which takes place slowly over the entire range between the first and second high temperature endotherms with a gradual change in the spacings of the *x*-ray diffraction lines but no loss in the intensity of the pattern. Crystal water is essentially completely removed by the end of the second high temperature endotherm. Water loss during this endotherm may well contribute to the thermal reaction but cannot fully account for it. During this endotherm the intensity of the *x*-ray powder pattern decreases rapidly indicating breakdown of the crystal structure and the formation of a disordered state which results in an increase in entropy. Thermodynamically the effect of increased entropy can be expressed by the general equation of third law, $dS = q/T$, where S is entropy; T, temperature; and q heat absorbed. Thus q increases with S and manifests itself as an endothermal reaction accounting for the second high temperature endotherm. The exothermic reaction which follows is attributed to the formation of spinel and partial return to an ordered state. Silica is not present as a crystalline phase in this temperature range.

ORIGIN OF LISTWÄNITES OF THE CAUCASUS AND URALS

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Listwänites are widely distributed in Eastern Europe and particularly in the Caucasus and Urals. At least three different types of listwänites have been observed: serpentinitic, gabbroic and porphritic. Here I shall present some findings concerning the serpentinitic listwänites investigated by me in those regions.

The term "serpentinitic listwänite" identifies a metamorphic rock composed chiefly of different carbonates (ankerite, dolomite, breunnerite, magnesite, calcite) with varying quantities of quartz (from 25 30% to 70%) and accessories, such as antigorite, chlorite, magnetite, chromite and others.

Contrary to the views of academician Karpinsky and Professor Lodschnikov, I have shown that the serpentinitic listwänites are products of hydrothermal alterations of serpentinites and that they cannot be considered as metamorphosed limestones as was originally supposed by those two scientists. Listwänites occur in the contact zones of the serpentinite bodies and its formation and composition appear to be independent of the chemical composition of the enveloping country rock. These rocks are formed as a result of a replacement of the serpentinitic rocks by silicon-bearing carbonic hydrothermal solutions.

Deposition of commercial quantities of gold (on berezites) during this replacement process was observed as an interesting side-light. In fact, listwänites were frequently used as ore-indicators by gold prospectors.

The comparatively high content of CaO and SiO_2 on the listwänites is a good illustration of Larsen's polarity law.

RECENT RESULTS FROM EARTHQUAKE SURFACE WAVE INVESTIGATIONS

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The dispersion and absorption of mantle Rayleigh waves with periods of 70–400 sec. have been investigated. The observed dispersion may be accounted for by the known increase of shear wave velocity with depth in the mantle. A value of .000018 km^{-1} was deduced for the absorption coefficient of waves with periods of 230 sec. The internal friction in the upper mantle corresponding to this absorption is $1/Q=590\times10^{-5}$, a value surprisingly similar to that reported from audio frequency vibration studies on crystalline rocks at atmospheric pressure.

Investigation of the short period surface shear wave Lg, which propagates a cross continental path only with a velocity of 3.51 km/sec. suggests that the large amplitudes and long duration of this phase can only be accounted for by wave guide transmission through a remarkably homogeneous sialic plate which ends abruptly at the continental margins.

Investigation of Rayleigh waves traversing oceanic paths with periods from 16–40 sec. reveals that crustal structure beneath the Atlantic and Pacific oceans is characterized by the absence of any significant thickness of granitic rock and the occurrence of the Mohorovician continuity at a depth of about 10 km below the sea surface.

RELATION BETWEEN IONIC SUBSTITUTION AND SWELLING IN MONTMORILLONITES

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The swelling of samples of Na-saturated montmorillonite from 12 localities was determined after separation of the clays from associated minerals. The degrees of swelling, which ranged between 21 and 66 ml per gram of sample, showed no correlation with cation-exchange capacity or with tetrahedral charge; but it did show a correlation with octahedral substitution, in the direction of decrease in swelling with increase in octahedral substitution. Substituted Fe^{+2} has a greater depressing effect on swelling than Fe^{+3}, which in turn appears to have about the same depressing effect as Mg^{+2}.

Swelling, being an osmotic property of montmorillonite, is affected by degree of ionization. The more complete the ionization, the greater the number of silicate sheets carrying a negative charge, the greater the tendency for the sheets to repel one another, and the greater the swelling; and vice versa. It is suggested that just as substitutions in acetic acid affect the strength and degree of ionization of the acid, so substitutions in the montmorillonite structure affect its degree of ionization and its swelling. The substitution of the more electropositive Fe^{+3} and Mg^{+2} ions for Al^{+3} would tend to alter the surface energy of the montmorillonite structure with the result that, when the mineral is dispersed in water, the degree of ionization is less and the consequent swelling is less.

PETROGRAPHIC EVIDENCE INDICATING POSSIBLE PATHS ALONG WHICH MATTER HAS MOVED IN AND AROUND QUARTZ AND FELDSPARS IN SOME METAMORPHIC ROCKS

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By appropriately adjusting the illumination in an ordinary petrographic microscope, a considerable amount of detail, commonly overlooked, can be seen. Great relief differences can be produced enabling one to distinguish with ease quartz and K-feldspars by inspection in plain light. "Channels," outlined by a Becke line, occur in and around the edges of certain minerals. Zoisite, clinozoisite, sericite, and other minerals develop along the channels and in plagioclases where the channels leave quartz grains. The color of biotite has changed where the channels cross it indicating compositional changes. The materials, added and subtracted, are thought to have moved along the channels.

No birefringence difference exists in quartz between the material in the channels (which has the lower index of refraction) and that on either side. This has tentatively been interpreted to mean that "chunks" of quartz (fragments of the lattice or particles of colloid size) have been removed and "porous," channel zones produced. A possible mechanism for this process is mentioned. H_2O molecules, H^+ or OH^- ions now occupying these zones produce the lower index of refraction noted in the channels.

Photomicrographs illustrating these phenomena and laboratory experimental results are presented to support the above conclusions.

PETROLOGY OF THE OLIVINE AMPHIBOLITE OF PARRY SOUND, ONTARIO, CANADA

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An island, approximately 260 by 150 feet, made up of olivine amphibolite is located in Blackstone Lake 15 miles southeast of Parry Sound. The rocks of the western part of the island show curvi-columnar structure.

The amphibolite consists of cummingtonite, olivine and serpentine with small amounts of spinel, magnetite and chlorite; chlorite is locally abundant. A colorless amphibole, perhaps gedrite, forms coronas between olivine and the amphibole.

With the exception of olivine the minerals of the amphibolite are of secondary origin and have probably formed from olivine and pyroxene, while serpentine and chlorite also formed from amphibole. The chemical composition of the rock comes closest to that of a picrite.

Pegmatite dikes are abundant in the area and diorite dikes cut the amphibolite. The formation of the olivine amphibolite from an original peridotite is probably due to metasomatic metamorphism caused by the dikes. The amphibole is believed to have tormed at the expense of pyroxene; Mg, Fe and Al ions in excess of those required for amphibole formation aided the formation of spinel. Near the dikes the rock is essentially a spinel amphibolite, and olivine is absent; the introduction of Al appears to have converted the olivine into amphibole and spinel.

DEUTERIUM CONTENT OF NATURAL WATERS

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A mass spectrometric method for the accurate determination of the hydrogen deuterium ratio has been developed. It is possible to determine the ratio of $\pm 0.10\%$ using material of "normal abundance" i.e. 1 part D in 6,700 parts H. Samples as small as 0.1 mg. H_2 (0.001 ml. H_2O) can be run. Natural evaporation and condensation that have been shown to fractionate the oxygen isotopes, also fractionate the hydrogen isotopes. The ratio of these two fractionation processes is equal to the ratio of the ratios of the H_2O/HDO and the H_2O^{16}/H_2O^{18} vapor pressures.

Ocean waters vary from $+3.25$ to $+5.65$ mole % deuterium, while fresh waters of the U. S. range from -10.1 to $+4.3$ mole % deuterium, both relative to an arbitrary standard. An explanation of these variations based upon the measured fractionation coefficient between liquid water and vapor will be given.

STUDIES OF URANIUM MINERALS (XI): GUMMITE

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The name gummite has been applied in a generic sense to designate fine-grained, pseudomorphic alteration products of uraninite whose true identity is unknown. It is similar in usage to limonite and wad. Study of a suite of gummite pseudomorphs from 35 localities has shown that the true identity of the mineral phases present can be established in most instances by *x*-ray powder methods. The alteration of the parent uraninite is in general sequential: (1) oxidation of U^4 in large part of U^6 with preservation of the uraninite structure, color usually black, brownish black or greenish black; (2) complete oxidation to U^6 with the formation of new anhydrous or slightly hydrated oxide phases, chiefly clarkeite and "green uraninite," color brown to green; (3) formation of new, relatively highly hydrated oxide phases, chiefly fourmarierite and "Mineral X" of Palache and Berman (1933), rarely schoepite, color yellow, orange or red; (4) formation of uranium silicates, chiefly uranophane and kasolite, less commonly beta-uranophane, rarely soddyite, color greenish yellow to straw yellow. The minerals of (2), (3) and (4) comprise gummite proper; mixtures are common. Clarkeite is a valid species, apparently one of the polymorphs of UO_3. "Tetragonal uraninite" and U_3O_8 were not observed in the course of the work.

STUDIES OF URANIUM MINERALS (XII): BILLIETITE AND BECQUERELITE

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Billietite ($BaO \cdot 6UO_3 \cdot 11H_2O$), originally described from Katanga by Vaes (1949), is a valid mineral species isostructural with becquerelite ("$2UO_3 \cdot 3H_2O$"). New chemical analyses and Weissenberg *x*-ray studies on two specimens of billietite from Katanga gave: (1) $a_0 = 12.04$ Å, $b_0 = 14.24$, $c_0 = 15.06$; $nX = 1.730$, $nY = 1.810$, $nZ = 1.815$; $BaO = 6.88$, $UO_3 = 82.76$, $H_2O = 8.97$, $SiO_2 = 0.76$, $CaO = 0.30$, total$=99.76$; S.G.$=5.57$ (meas.), 5.33 (calc.); (2) $a_0 = 12.08$ Å, $b_0 = 13.98$, $c_0 = 15.06$; $nX = 1.725$, $nY = 1.780$, $nZ = 1.790$; $BaO = 7.41$, $UO_3 = 84.39$, $H_2O = 8.68$, total$=100.48$; S.G.$=5.15$ (meas.), 5.40 (calc.). Billietite is orthothombic (2V(—)$\sim$35°; X pale yellow, Y and Z deep golden yellow. There are $4(BaO \cdot 6UO_3 \cdot 11H_2O)$ per unit cell. The unit cell differs from that of Brasseur (1949) in that the *a*- and *b*-axes are interchanged and the *b*-axis is doubled. Brassuer has $a_0 = 7.11$ Å, $b_0 = 12.08$, $c_0 = 15.09$; $nX = 1.730$, $nY = 1.822$, $nZ = 1.829$; $BaO = 6.95$, $UO_3 = 83.36$, ign. loss$=10.49$, total$=101.30$.

A new analysis and *x*-ray study of becquerelite from Katanga gave: $a_0 = 12.45$ Å, $b_0 = 13.92$, $c_0 = 15.09$; $nX = 1.770$, $nY = 1.805$, $nZ = 1.820$; $UO_3 = 89.53$, $H_2O = 8.95$, $SiO_2 = 1.82$, total 100.30; S.G.$=5.3$ (meas.), 5.56 (calc.). Becquerelite is orthorhombic, 2V(−) 30°; X yellow, Y and Z deep golden yellow. There are 14 ($2UO_3 \cdot 3H_2O$) in the unit cell.

The near coincidence in dimensions and intensities of the single crystal and powder photographs of billietite and becquerelite indicate that these minerals are isostructural. From the new analysis a formula ($7UO_3 \cdot 11H_2O$, with S.G. calc.$=5.60$) for becquerelite can be written analogous to that of billietite. The unit cell contains four formula units. A substitution of one Ba for one U^6 is presumed. If the role of water in the structure were understood, it might be possible to explain the necessary valence compensation for such a substitution by the O to (OH) ratio in each mineral.

EXTENSION OF THE "DIFFERENCE PATTERSON" TO FACILITATE THE SOLUTION OF ORDER-DISORDER PROBLEMS IN MINERALOGY

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Great difficulty is encountered in attempts to locate the specific metal positions from among the symmetrically distributed possible positions in those minerals based upon close packed sulfur atoms (chalcopyrite, bornite, stannite, etc.). Because of the identity of many interatomic distance vectors the Patterson projections, or sections, of such structures cannot be readily solved, even by modern techniques.

If it can be shown, or assumed, that the non-superstructure x-ray reflections do not change in intensity when these structures are disordered then Patterson projections or sections made using the superstructure reflections alone represent vector maps where all vectors are suppressed except those between like atoms that change position during the disordering process. Positive peaks represent vectors between sites that have either both gained or both lost in atomic number through disordering. Negative peaks take on a significance; they represent vectors between two sites one having gained and the other having lost in atomic number, hence represent the vectors between unlike atoms that change position during disordering. Chalcopyrite is used as a known example to show the advantages and shortcomings of the method. The method is being applied to the bornite problem.

URANIUM DEPOSITS IN THE BLACK HAWK DISTRICT, GRANT COUNTY NEW MEXICO

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Pitchblende occurs in association with native silver and with nickel and cobalt arsenides and sulfides in the old Black Hawk district, 14 miles west of Silver City, New Mexico. The geologic setting is similar to that of the high grade uranium deposits at Great Bear Lake, Canada, and at Joachimsthal, Czechoslovakia; and the mineralogical associations in the three localities are almost identical. In geologic setting and in mineralogical associations the Black Hawk deposits are unique in the United States. The Black Hawk district produced high-grade silver from 1881 to 1893, but the mines have been idle since that time and are at present inaccessible.

The ore is in fissures with a carbonate gangue in pre-Cambrian quartz diorite gneiss. The veins trend north to northeast; and the deposits that contain radioactive minerals form a northeast-trending belt about 2 miles long. This belt parallels the elongation of a nearby monzonite porphyry stock. Most of the major deposits are along faults in the gneiss, adjacent to where the faults cut monzonite porphyry dikes or apophyses of the main stock. Diamond drilling to test the area for commercial deposits of uranium, nickel, and cobalt is in progress.

MICROCLINE-SANIDINE STABILITY RELATIONS

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Natural potash feldspars that are crystallographically intermediate between microcline and sanidine are described. It is not yet known whether the intermediate states represent stable configurations of a disordering transition as a function of temperature, or whether they are non-equilibrium states.

Conversion of microcline to monoclinic $KAlSi_3O_8$ can be induced by hydrothermal treatment, but the reverse (i.e., "synthesis" of microcline) has not been accomplished and probably will not take place in times available for experimentation. *X*-ray patterns show that a small percentage of finely powdered microcline is converted when held several days at 550° C. (8,000 psi H_2O pressure). Conversion is apparently complete in this time at 800° C.

Hydrothermal treatment at 800° C. of small microcline fragments used for single-crystal *x*-ray analysis show the development of tiny oriented overgrowths of monoclinic feldspar on the microcline. This phenomenon, coupled with the fact that fine powder is converted to the monoclinic form relatively rapidly whereas the small fragments (0.2 mm.) take very much longer indicates that the conversion is brought about by surface recrystallization abetted by the hydrothermal environment. The fact that hydrothermal recrystallization to a monoclinic phase (that may be metastable at the temperature of crystallization) probably takes place in considerably less time than does stable internal readjustment in microcline, makes experimental interpretation of the transition temperature or temperature range somewhat uncertain.

THEORETICAL CONSIDERATIONS IN GAMMA-RAY WELL LOGGING

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Gamma-ray logs are useful in determining subsurface stratigraphy in uncased and cased oil field drill holes. Logs of diamond drill holes may also be used to evaluate uranium reserves. A theoretical treatment of the problem indicates practical conclusions that have been overlooked in the usual empirical applications. Statistical fluctuations in the emission of gamma rays limit the resolution and logging speed. The use of scintillation counters affords some improvement. The theory allows corrections to be made for the increased apparent thickness of beds and the depth displacement caused by the finite length of the detector and the time constant of the circuitry. The effect of multiple strings of casing, eccentric location of detector, shielding around the detector and radioactivity of mud and mud cake can be evaluated individually. The theory also is useful in evaluating the efficiency of radioactive tracers, the depth of penetration of radioactive bullets, and the radial weighting function in neutron-gamma logs. Limitations in the present treatment are: (1) a purely exponential and inverse square attenuation is assumed, (2) a uniform horizontal distribution of radioactive material in the formations is assumed, (3) the dip is assumed zero and (4) the counting rate of the detector is assumed to be directly proportional to the gamma flux on the drill-hole axis.

CHEMICAL VARIATIONS IN THE PRE-TERTIARY ROCKS OF CORNUCOPIA, OREGON

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The Pre-Tertiary rocks of the Cornucopia area, in the southeastern part of the rugged Wallow Mountains of northeastern Oregon, are chiefly metamorphosed sedimentary rocks of Paleozoic and Mesozoic age, and granitic rocks of varying composition. Hornfelses derived from argillaceous sedimentary rocks are dominantly of mesozonal grade.

The contacts of the granitic rocks with the metamorphic rocks are gradational. For example a granitic mass 232 feet wide cut by a lower adit, has transitional zones 30 and 38 feet respectively. Chemical variations across this zone include about 10 per cent increase in SiO_2 in the granitic rock with a corresponding decrease of nearly 5 per cent in Fe_2O_3+FeO

and over 2 per cent MgO but with little change in CaO and Na_2O. Crystalloblastic textures prevail in the granitic rock as well as in the transitional zones.

From the 31 analyses afforded by a G.S.A. grant, several variation diagrams have been made and these diagrams are quite different from those of magmatic rocks. It is the opinion of the writer that the chemical variations can best be explained by the hypothesis of granitization rather than magmatic differentiation.

STABILITY OF MAGNESIAN CALCITES

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The geologic stability at earth surface conditions of calcite solid solutions with at least 1 or 2 mole per cent $MgCO_3$ is suggested by the occurrence of such material in the Quaternary and Tertiary of Florida, and elsewhere.

However, it is also a commonly held view that the difference of size of Ca^{++} and Mg^{++} precludes the formation of stable magnesian calcites. In order to evaluate these matters, the stability of the hard parts of certain present-day algae and echnoids containing as much as 20 mole per cent $MgCO_3$ in solid solution has been studied. Ten day runs at 300° C. in the presence of water expel little or none of the Mg, but similar runs at 400° C. produce a pure calcite and brucite. The presence of moderate CO_2 pressure appears to inhibit the breakdown of the solid solution.

Magnesian calcites containing at least 20 mole per cent $MgCO_3$ in solid solution have been produced synthetically both at room temperature and at elevated temperatures and pressures, but these products may be metastable. In the former case, in particular, the conditions under which crystal growth begins are critical; a high concentration of NaCl decreases the rate of growth and favors greater content of Mg in the carbonate solid solution.

SYNTHESES OF URANIUM MINERALS AT ROOM AND ELEVATED TEMPERATURES*

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Syntheses of uranium-minerals is probably the best way of solving some of the problems accompanying the mineralogy of the several scores of natural uranium-compounds. Uranyl minerals, that is, so called secondary ones, were synthesized at room and elevated temperatures in solutions that usually had a terminal pH of 1 to 5. Zippeite ($5UO_3 \cdot 3SO_3 \cdot 10H_2O$) and becquerelite ($5UO_3 \cdot 10\text{–}12H_2O$) were made at room temperature. Fourmarierite ($3PbO \cdot 6UO_3 \cdot 5H_2O$) formed easily at 100° C. Curite ($3PbO \cdot 7UO_3 \cdot 4.5H_2O$), kaselite ($PbO \cdot UO_3 \cdot SiO_2 \cdot H_2O$) and soddyite ($5UO_3 \cdot 2SiO_2 \cdot 6H_2O$) readily formed at 180° C. and higher. The silica in kasolite and soddyite came from the glass containers of the experiments. Clarkeite ($UO_3 \cdot nH_2O$) is the only mineral recognized in solutions with a pH higher than 7 at 258° C. In many experiments one or two unknown compounds were found. Pitchblende (uraninite) was made at 50° to 212° C. by reduction of uranyl ions with H_2S or ferrous ions. Reactions below 50° C. are extremely slow. In some experiments at 212° C. hematite was a by-product of the pitchblende which could explain the common hematitic staining in hydrothermal veins. Organic liquids are able to reduce uranyl ions to pitchblende at room temperature. Whether these organic reactions take place in the complete absence of light is not yet known. That they occur in ultraviolet radiation has been known for 50 years.

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* This investigation is a part of a contract with the Raw Materials Offices of the Atomic Energy Commission and is published with their permission.

INTRUSIVE TUFF-BRECCIA

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Andesitic tuff-breccia of intrusive origin is widely exposed over an area of at least 100 square miles in northwest Wyoming in the east-central part of the Absaroka Range, a dissected volcanic plateau consisting mainly of Eocene deposits. The tuff-breccia occurs as one or more essentially horizontal sheets, an unusual feature for intrusive pyroclastic rock. The intrusive sheet(s) range in thickness from a maximum of about 650 feet to a minimum of 50 feet or less. Included blocks from the overlying beds, some as much as several hundred feet in diameter, are a characteristic feature of the tuff-breccia. There is contact alteration of both these inclusions and the country rock beds.

Stratigraphic evidence suggests that the tuff-breccia was emplaced at relatively shallow depth (probably between 600 and 1900 feet); locally it probably reached the surface. The beds into which the pyroclastic rock was intruded were probably poorly consolidated and water-saturated. The occurrence of foreign material unlike nearby country rock and beds of the underlying sequence suggests underground lateral movement of several miles.

The temperature at the time of emplacement is unknown, although petrographic evidence suggests that the tuff-breccia did not contain liquid magma at the time of emplacement. Some heat (and volatiles) was probably required, however, for the contact alteration, characterized by chlorite(?) and hardening (silicification ?).

No satisfactory explanation can now be given for the mode of origin and emplacement, but evidence seems to suggest the relatively non-explosive subcrustal formation of tuff-breccia and its injection under pressure.

CHEMICAL DIFFERENTIATION OF MULTI-GENERATION PEGMATITE MINERALS

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Differentiated pegmatities, consisting of zones or combined zones and younger hydrothermal units, can contain more than one generation of a mineral species. Compositional variation of successive generations of a species is systematic and is a function of paragenetic position. Decreasing relative age of formation is accompanied, in plagioclase, by increasing Na_2O; in microcline by decreasing amounts of sodic plagioclase as exsolution perthite and by increasing Rb_2O. Alkalies, except K_2O, also increase in lepidolite, tourmaline, beryl and some phosphates, with younger generations of each. In garnet, tourmaline, columbite-tantalite, triphylite-lithiophilite, muscovite and probably lepidolite the Fe:Mn ratio decreases with decreasing relative age. Likewise in columbite-tantalite Ta_2O_5 increases in younger members. Other multigeneration species that reveal compositional differences are amblygonite (in the F-OH ratio and in Na_2O), apatite (in the F-OH ratio and MnO), cassiterite, biotite, spodumene, quartz and topaz. The small-scale, sequential changes in the composition of succeeding generations of pegmatite minerals reflect, in a general way, the large-scale variations in the magmatic differentiation series, gabbro to granite, and thus constitute an argument for the origin of structurally complex pegmatites through fractional crystallization.

METASOMATISM BY APOPHYSAL DIKES IN THE CRAWFORD NOTCH REGION, NEW HAMPSHIRE

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On the northwest side of Crawford Notch in the White Mountains of New Hampshire, a large ring-dike of Albany porphyritic quartz syenite fingers out into many apophysal dikes where it meets a body of fine-grained gneiss of the Littleton formation. The rock in the ring-dike proper is porphyritic quartz syenite which is composed of approximately equal amounts of euhedral phenocrysts of feldspar 1 cm. across, and of 0.5 mm. grained groundmass. The rock of the apophysal dikes ranges from quartz syenite porphyry to granite porphyry. The groundmass is dark-gray to black, aphanitic, and composes two-thirds of the rock. The phenocrysts are similar to those in the porphyritic quartz syenite.

The apophysal dikes are enclosed in thick to thin bedded, fine-grained biotite gneiss and hornfels which are locally andalusite bearing. The dikes range from a few feet to 300 feet wide and from a few tens of feet to several hundred feet long. Most of the apophysal dikes are bordered by transitional zones ranging from a few inches to 25 feet wide which grade into fine-grained gneiss and hornfels. The most striking feature of the border zones is the occurrence in gneissic matrix of abundant feldspar "phenocrysts" like those of the dikes proper. The groundmass of the transitional zone is chiefly fine-grained gneissic material with abundant added tourmaline and intermixed constituents of the groundmass of the dikes.

It is concluded that significant amounts of rock-forming minerals as well as "mineralization products" such as tourmaline, sulfides and silicification can be produced by metasomatic action of magmatic dikes even as fine-grained as porphyries.

STRUCTURE AND METASOMATISM OF THE BELT SERIES NORTHWEST OF THE IDAHO BATHOLITH*

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The older rocks around the northwestern corner of the Idaho batholith belong to the pre-Cambrian Belt series.

The area is structurally complex, situated at the junction of two arcuate segments of Nevadan folding and also on the western border of the Laramide orogenic belt. The major fold axis parallels the trend of the northern arcuate segment of Nevadan folding (N. 70°–80° W.). The lineation and axis of small folds parallel the trend of the southern arcuate segment (N. 50° E.). Both sets of folds are locally overturned to the south and are accompanied by low-angle thrusts. The gentle north trending folds may be of Laramide age. The block faulting continued to Miocene time.

The main folding was accompanied by small intrusions of gabbro, diorite, and tonalite, intense metamorphism, and metasomatic introduction of hornblende, biotite, andesine, and accessories. Locally, a second period of metasomatism obliterated the folded structures and is interpreted as having caused the formation of monomineralic rocks like hornblendites and anorthosites. This second period of metasomatism may be contemporaneous with the emplacement of the quartz dioritic border zone of the batholith. Small intrusions of tonalite west of the batholith and the rocks of the quartz monzonite series in the center and north of the batholith were emplaced after the second period of metasomatism.

Chemical analyses show that mainly silica was expelled and Fe, Mg, Al, Ca, and Ti were introduced during the metasomatism.

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* Publication authorized by the Director, U. S. Geological Survey.

RADIATION DAMAGE IN ZIRCON: A POSSIBLE AGE METHOD

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The relationship has been studied between crystal structure damage and total alpha particle irradiation in zircon. The total irradiation was based on estimated or measured age times alpha activity, and the damage was determined by the change of the x-ray diffraction angle from the (112) plane as measured to .01° by x-ray spectrometer. The angle was found to be 35.635° (2θ) for completely undamaged material and it approached 35.1° asymptotically as the zircon had suffered increasing dosages of alpha irradiation and become metamict. The difference between the measured angle and 35.635° was found to equal $0.535e^{-2.31 \times 10^{-16}\alpha}$ degrees, where $\alpha =$ total irradiation in alphas/mg. The results have been interpreted to indicate that 4.5×10^3 atoms are displaced per alpha disintegration and that the annealing rate is essentially zero at earth's surface temperature. The process appears to be useable for age measurement but generally limited to zircons of less than 500 alphas/mg./hr. activity.

It was further found that the zircon became "half-metamict" after a radioactive energy release of 685,000 cal./gm., indicating that only a fraction of a per cent of the energy is stored in the structure as disordered atoms, the remainder being lost as heat. Observations on the high annealing rate of common rock minerals show that the amount of energy capable of being stored in a rock by alpha damage is negligible in questions relating to petrogenesis and volcanism.

HEAT PRODUCTION IN BASALTS AND THEIR ORIGIN

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Progress is being made on the measurement of the radioactivity of crustal rocks (including potassium) for the determination of heat generation. Present averages show 2.2×10^{-6} cal./gm.yr. for the basalts of the Columbia Plateau, Oregon, 1.0×10^{-6} for the intrusive diabase dikes of the Canadian Shield area, and 1.0×10^{-6} for the Deccan basalts of India.

The objective is a clearer definition of conditions limiting the origin of certain basic volcanic rocks, particularly those of a tholeiitic flood type, those associated with stable Shield areas, and those of an oceanic type. At the present time it appears that the thermal gradient in large, low-lying areas of old crystalline rocks, even allowing for a low average radioactivity to the depth of the Mohorovicic discontinuity, will not permit the temperatures needed for the production of basic magmas at depths less than 100 km under equilibrium conductive conditions, and at the same time the existence of a layer of material of basaltic composition at these depths for any great time is not permitted on the basis of its heat generation.

It therefore seems necessary to think in terms of at least local, transient, thermal columns caused by convected material in the mantle. If the convected material is of a peridotitic composition it would be necessary that a gravitational separation of a basalt-like extract be effected at some time during the history of the column. This extract could not represent more than one part in 10^{-2} of the original mantle material.

STATISTICAL STUDY OF TEXTURE OF ENCHANTED ROCK PLUTON, SOUTHWESTERN LLANO COUNTY, TEXAS

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Enchanted Rock pluton, one of Stenzel's Town Mountain Series of plutons, covers 100 square miles, is of pear-shaped outline, and intrudes tightly folded Precambrian metamorphics.

Tectonic features within the mass and structural attitude of wallrocks indicate the northern third is phacolithic, but the remainder is probably discordant batholithic. The phacolithic part occupies a synclinal trough plunging 16–20° southeast.

Density of the granite varies 2.62–2.68. Biotite-rich schlieren vary 2.93–3.02. Xenoliths run commonly 2.71–2.74 and may strike 15–30° transverse to surrounding platy flow structure.

Seven traverses were run from center to margin of the pluton normal to platy flow structure. Number and dimensions of microcline phenocrysts were recorded for a 100 square centimeter area at each station. Chi-square values were calculated for their long dimension. On horizontal cross section crystals are rectangular. Brachy- and macropinacoids predominate.

Variation diagrams showing number of crystals per 100 square centimeters and average area of these crystals have characteristic curves. Chi-square values show only probability that deviations from the hypothesis of independence are caused by chance.

Characteristic curves indicate textural changes and aid in determining cooling history and structure of the pluton. Judging from chi-square values, probability that curves obtained are due to chance is small. Chi-square values are especially significant when comparing dimensions of crystals occurring in chill zones with those in which chilling was weak to absent. Where chilling occurred values are low; where crystallization proceeded uninterrupted values are high.

OCCURRENCE OF OPTICALLY NEGATIVE FORSTERITE IN THE SYSTEM Na_2O-MgO-Al_2O_3-SiO_2

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Biaxial negative forsterite was crystallized from melts containing Na_2O, MgO, Al_2O_3 and SiO_2. The crystals show inclined extinction up to 10° and prominent embayments. Optic axis interference figures indicate an optic angle lower than that reported in the literature for pure forsterite. *X*-ray diffraction patterns are identical with that reported in the literature for pure synthetic forsterite.

Although the ionic radii of both Na^+ and Al^{3+} differ radically from Mg^{2+}, the sum of the radii of Na^+ and $Al^{3+} = 2Mg^{2+}$. It is postulated that atomic substitution takes place on a limited scale to account for the change in optical properties and morphology of the crystals.

RATIO OF OXYGEN ISOTOPES IN QUARTZ OF CONTRASTED ORIGIN

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Measurements of the $O^{18}:O^{16}$ ratio in carbon dioxide produced by reaction of quartz with carbon in a vacuum furnace show that samples of quartz from a geological environment indicative of relatively low temperature of quartz growth, have a higher content of

O^{18} than do quartz samples formed at higher temperature. Differences are small but readily measurable with the mass spectrometer. Results indicate that the oxygen isotope ratio of quartz, taken in conjunction with other data, including data on the high-low inversion, may provide a useful clue to geologic temperatures in a relatively low range.

ALTERATION AND URANIUM MINERALIZATION, MARYSVALE, UTAH

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Alteration zones containing a sequence of clay minerals, silicification, alunitization or pyritization occur in a variety of rock types in an area some fifty square miles in extent at Marysvale, Utah. Two field seasons with an intervening period of laboratory study have been devoted to an analysis of the relationship between the alteration and uranium mineralization. More significant channels of alteration have been mapped.

The area has been mapped geologically in order to establish correlation between the alteration and rock types. It has been found desirable to subdivide the major geologic units, such as the Bullion Canyon volcanic series, the Mount Belknap volcanics and the intrusives for greater clarity in structural interpretation, interpretation of the alteration and the delineation of host rocks presumably favorable to the deposition of uranium.

Alteration appears to diminish in the Mount Belknap red facies. Minor bleached areas occur along faults and shear zones. These are characterized by the development of adularia but contain few clay minerals. Such zones traced into bordering rocks of the Bullion Canyon volcanic series or intrusives may show enlargement and the development of argillic alteration.

Uranium bearing veins contain both oxidized uranium minerals and uraninite. Fluorite is a common associate becoming more darkly colored near ore zones. Jordisite and ilsemannite are associated vein minerals. Montmorillonite, hydromica and kaolinite are common clay mineral constituents of the uranium vein walls. Localized hematitic alteration may be observed. Biotite may develop as an alteration mineral. Brecciated thin quartz veins, pyritic and chloritic zones and occasionally magnetite zones occur in the uraninite area.

Uranium minerals have been found in several units of the Bullion Canyon series accompanied by zones of argillic alteration. The principal veins found to date, however, are in quartz monzonite or fine-grained granite. The Mount Belknap has not yet yielded more than migratory secondary mineralization along fractures. Alunitic zones are generally free from uranium, but uranium mineralization is occasionally found nearby.

Conclusions presented constitute a progress report on the Marysvale Alteration Study jointly sponsored by the Division of Raw Materials, Atomic Energy Commission and the Department of Geology, Columbia University.

CARBON 14 MEASUREMENTS ON GEOLOGICAL SAMPLES

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Carbon 14 age measurements on various types of geological samples which have been made during the last year will be reported. Wood specimens from cores taken in the Mississippi delta region show that the last major sea level lowering (on the order of 400 feet) occurred more than 30,000 years ago and that the rise has been more or less continuous since then. The time of a certain sea level in the Mississippi delta region is very close to the same level on Bermuda. A survey of the carbon 14 content of the atmosphere indicates that it is essentially constant and corresponds closely to that for living wood. The carbon 14 concentration in ocean carbonate is much less for deep water than for surface water suggesting that the rate of ocean turnover is on the order of thousands of years. There is evidence for a variation in the carbon 14 concentration of the surface ocean water. This is important in the evaluation of ages derived from shellmaterial.

PETROGENESIS OF A GABBRO-GRANOPHYRE COMPLEX IN NORTHERN WISCONSIN

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A gabbro-granophyre complex is exposed in northern Wisconsin on the steeply dipping south limb of the Lake Superior geosyncline. The complex consists of two distinct belts of middle Keweenawan gabbro and granophyre split by a narrow belt of lower Keweenawan extrusives. Younger basic dikes cut the area.

The gabbro was intruded along zones of weakness developed between certain flow units during the formation of the geosyncline. In the gabbro two apparently unrelated structures are found. The first, fluxion structure, is believed to be due to shear or flow as the mass crystallized. The second, rhythmic layering, appears to be related to crystal settling. Gabbroic mineral compositions reveal simultaneous soda enrichment in plagioclase and iron enrichment in coexisting pyroxenes and olivines.

Between the gabbro and granophyre is a narrow zone of "intermediate rock" which exhibits peculiar mineralogical and textural characters. Fourteen features of this zone indicate that it originated as an alteration product of gabbro through contact action of a granophyric melt. For the granophyre, criteria of a replacement origin are inconclusive. Eleven facts strongly support a magmatic origin. The granitic magma may have developed through partial refusion in the lower reaches of the geosyncline or through fractional crystallization at depth. The resulting granophyric texture is unrelated to eutectic crystallization.

As a result of the contact action of granophyre magma on solidified gabbro, geochemical culminations of TiO_2, FeO, MnO, and P_2O_5 occur in the "intermediate zone." Variations in isotopic compositions of oxygen also resulted from this contact metasomatism.

RELATION OF PITCHBLENDE DEPOSITS TO HYPOGENE ZONING IN THE FRONT RANGE MINERAL BELT, COLORADO

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The distribution of pitchblende deposits in several Front Range mining districts shows a distinctive relation to the pattern of hypogene zoning. The deposits appear to be restricted to an intermediate or transitional zone between central areas of pyritic gold veins and peripheral areas of silver lead-zinc veins. At present we do not know whether pitchblende is more common in inner or the outer part of the intermediate zone. In at least one district (Central City) significant quantities of copper in the form of chalcopyrite have been produced from the intermediate zone. The position of uranium in an intermediate zone between gold and zinc was tentatively implied by W. H. Emmons (1926) in his generalized treatment of hypogene zoning.

At least seven local geologic factors must be considered in interpreting the zonal distribution of Front Range pitchblende deposits. These factors are: bostonite intrusives, fault patterns, multiple stages of uranium mineralization, differences between paragenetic sequence and position of minerals in successive zones, correlation of pitchblende with copper sulfides (and sulfosalts?), scale and complexity of zoning in plan and in section, and effective depth of mineralization. The significance of some of these factors can already be evaluated.

The concept of the position of pitchblende in hypogene zoning is developed to stimulate and guide the systematic search for pitchblende deposits. Though the zonal position of pitchblende deposits is best known for areas of complex base-metal mineralization dominantly of mesothermal (possibly xenothermal) character, knowledge of the zonal position of uranium deposits in widely different geologic environments might be similarly useful in a search for ore.

RELATIONSHIP BETWEEN POLYMORPHISM AND COMPOSITION IN THE MUSCOVITE-LEPIDOLITE SERIES

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A determination of the crystal structures of 500 muscovites and lepidolites, approximately 50 of which are of known composition has demonstrated the existence of a complete series between muscovite and lepidolite. The polymorphic variation in this series can be correlated with chemical composition, particularly with Li_2O content. A new variant of the muscovite structure has been discovered for which the term lithium muscovite is proposed. This modification crystallizes with a structure very similar to that of normal muscovite and apparently links heptaphyllite muscovite with octophyllite lepidolite. Micas with 3.2–4.0% Li_2O are generally characterized by poor crystal development and anomalous optical properties owing to their transitional structures. Lepidolites with 4.0–5.1% Li_2O generally crystallize as the 6-layer monoclinic polymorph whereas those with more than 5.1% Li_2O have usually crystallized as the 1-layer monoclinic polymorph. Rhombohedral lepidolite is commonly associated with the 1-layer monoclinic polymorph, and both have essentially identical compositions. Twinning may be responsible for the rhombohedral form. Polymorphic modifications of lepidolite may occur across sheets and also along the c axis directions in single books.

MINERALS OF THE SAPUCAIA PEGMATITE MINE: FAHEYITE, A NEW BERYLLIUM PHOSPHATE*

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Faheyite, a new mineral from the Sapucaia pegmatite mine, Minas Gerais, Brazil, has the composition $(Mn, Mg, Na)Be_2Fe_2(PO_4)_4 \cdot 6H_2O$. It occurs as white, bluish-white, or brownish-white minute fibers that coat minerals in vugs. Evidences for a late-stage phosphate mineralization include the occurrence of faheyite as tufts of fibers on altered muscovite, as flat rosettes between muscovite sheets, and as fiber-coating on euhedral crystals of quartz and variscite. The variscite crystals are unusual in that they are raspberry red and contain up to 2.46 per cent MnO. Crystals of roscherite, childrenite, amblygonite, and apatite and botryoidal masses of frondelite also occur in the vugs.

The powder pattern has been completely indexed on the basis of a primitive hexagonal cell with dimensions $a_0 = 9.43$ Å, $c_0 = 16.00$ Å; the cell contents are $Z = 3$. The powder film was compared with a rotation picture of a fiber taken in the powder camera, as an aid to indexing the powder film. X-ray powder data show strong lines at 5.72, 7.28, 3.243, 3.087, 3.031, and 3.958 Å. The specific gravity is 2.660 (obs.) and 2.670 (calc.). The indices of refraction are $\omega 1.631$ and $\epsilon 1.652$.

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* Publication authorized by the Director, U. S. Geological Survey.

The analysis shows the following percentages: insoluble 9.44, P_2O_5 38.11, Fe_2O_3 21.42, Al_2O_3 0.10, Mn_2O_3 none, FeO none, BeO 7.26, MnO 5.99, MgO 1.14, Na_2O 0.84, K_2O trace, F trace, H_2O 14.90, total 99.20.

The mineral is named in honor of Joseph J. Fahey, geochemist of the U. S. Geological Survey.

NATIVE LEAD, PRESIDIO COUNTY, TEXAS

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Approximately 3,000 pounds of ore containing native lead was recovered from three small tabular ore bodies in the western part of the Shafter District, Presidio County, Texas. Two of the ore bodies were vein-like in limestone with clay gouge walls. The metallic lead was localized at the centers of the veins. The third ore body was flat lying disk-shaped about 4 feet thick with a surface area of 1,600 square feet. The native lead occurred along a part of the margin of the ore body in a zone about 9 inches thick separated from the wall rock by clay gouge.

The ore generally was argentiferous steel galena partly or completely altered to cerussite and anglesite with minor minerals including wulfenite, litharge (?), quartz and limonite. The native lead occurred as threads, sheets to one-half inch thick and pods up to 4×7×2.5 inches.

The native lead appears to have been derived from the secondary minerals. Microscopic, chemical and spectrographic data are presented, and the paragenesis of the mineral suite discussed.

GEOLOGY OF THE GYPSUM DEPOSITS NEAR SUN CITY, BARBER COUNTY, KANSAS

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An area of approximately four square miles, in the vicinity of the Pioneer Mine, Sun City, Kansas, was mapped to ascertain the relationship between the Medicine Lodge gypsum and anhydrite. Samples were studied petrographically.

The Medicine Lodge gypsum is lower Permian in age and forms the cap rock in the area. The regional dip is about 11 feet per mile to the southwest. Structural contours on the base of the gypsum show local irregularities which may represent initial dip.

Anhydrite is not generally present on surface outcrop, but is exposed in the mine. It occurs as lenticular deposits about 10 feet above the base of the gypsum. The thickness of the gypsum is about 30 feet in the mine. Where anhydrite is absent, solution channels generally are present. These channels are zones of brecciation in which the gypsum is soft, porous, and sugary, and coated with some clay. Drilling has encountered water in these channels.

Petrographic studies show (1) gradation of anhydrite to gypsum; (2) gypsum fracture filling in the anhydrite; (3) absence of anhydrite in the upper and lower part of the gypsum; and (4) no evidence of distortion.

Hydration of this anhydrite may have taken place subsequent to the sedimentary anhydrite-gypsum deposition.

INSHORE MARINE MAGNETIC INVESTIGATIONS: THE AREA FROM NEW JERSEY TO CAPE COD, MASSACHUSETTS

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The total intensity of the earth's magnetic field was measured from a ship along various tracks from New Jersey to Cape Cod to a distance about 30 miles from shore. The measurements were made with a continuously recording airborne magnetometer which had been modified to be towed about 30 feet under water 200 or more feet astern the survey vessel. This survey shows that there is no magnetic anomaly associated with the inshore end of the Hudson Canyon. It found major anomalies of several hundred gammas magnitude covering large areas 10 miles east of Sandy Hook, N. J., south of Fire Island and SSW and NNE of Block Island. A non-anomalous region was found south of Narragansett Bay. The half widths of smaller anomalies are the order of fractions of a mile at the Cape Cod end of the area, while in the area south of Long Island they had half-widths of the order of miles. It is concluded that there is no major structure at the surface of the crystalline rock under the Hudson Canyon near shore. The Narragansett Basin is shown to extend south under water at least 10 miles. Several hypotheses concerning the subsurface crystalline rock in the area are proposed.

URANIUM ORE CONTROLS IN THE HAPPY JACK MINE AND VICINITY, WHITE CANYON, UTAH

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Channels within the Triassic Shinarump formation are thought to be the primary uranium ore control in White Canyon, southeastern Utah. This has been demonstrated in the Happy Jack Mine and on the Sunrise and Gonaway claims. In addition, there are important intra-channel controls; the most conspicuous being carbonaceous matter and lithologic changes.

Shinarump stream channels were first mapped in White Canyon during 1951, on the basis of outcrop mapping alone. Three of these channels, the Happy Jack, Sunrise and Gonaway were later explored and extended by drilling. Detailed geologic mapping of the Happy Jack Mine and drill hole information supplemented each other.

A structural contour map of the base of the Shinarump, corrected for regional dip, was the most important tool in locating ore. Each drill site was located only after the preceding diamond drill core information had been evaluated and placed on the contour map. Forty-five per cent of the holes drilled were situated in ore, resting on or near the bottom of channels. No ore was found in this area outside of channels. Correlations among diamond drill holes and outcrops reveal that the ore is characteristically in a porous carbonaceous sandstone on top of Shinarump siltstone.

Cross sections of the Happy Jack Mine show, in addition to the above features, horizontal bedding plane control for pitchblende and sulfide mineralization. There is no evidence for vertical control of primary ore.

SOME SPECIAL CRITERIA FOR GRANITIZATION

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In addition to textures, replacement features, structural and stratigraphic continuity with, and chemical relationships to country rocks, etc., the following criteria appear useful.

Early epidote often occurs in amphibolitic-derived granitized rocks; either inherited, or formed by early-metasomatic partial biotitization of hornblende. It antedates part of the plagioclase and all of the late K-feldspar. Its preservation indicates temperatures below those of any silicate melt. Microscopic relations between early epidote and feldspar greatly differ from those between feldspar and secondary epidote.

Early epidote and other minerals assign a considerable part of the granitized rocks to temperatures lower than those of the high-grade zone. Mapping of progressive zones leads to the concept of *external and internal zoning*, some isograds being outside and others inside the granitized body. At igneous contacts, all progressive zones are external.

Some relict minerals in granitized rocks are also chemical indicators of non-magmatic origin. *Al-silicates* have been previously discussed. Sometimes *lime-silicate minerals* survive. Chemical incompatibility of certain minerals with the bulk rock composition suggests granitization.

Many granitic rocks have unequivocally crystalloblastic, some indubitably igneous textures. *If crystalloblastic and igneous appearing features are combined, time analysis* is decisive. If igneous textures are older, an igneous rock has been altered autometamorphically or metamorphically. If crystalloblastic textures are older, a granitized rock has eventually developed certain igneous features. Final liquefication destroys all crystalloblastic and creates new igneous textures. Intrusive emplacement does not prove liquefication.

Chemical similarity of advanced granitized and igneous granitic rocks indicates close genetic connection, rather than two geologically unrelated origins. Probably the quantitatively apparently subordinate, true igneous granitic rocks mostly form by liquefication of granitized rocks. Granitization converges toward a chemical equilibrium in the crust.

PITCHBLENDE DEPOSIT AT THE CARIBOU MINE, BOULDER COUNTY, COLORADO

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Pitchblende occurs at the Caribou mine in a mesothermal carbonate vein that contains silver, lead, and zinc minerals. Unlike most of the large, well-known pitchblende deposits of the world, it does not contain nickel- and cobalt-bearing minerals. The deposit is believed to be unique in that the host rock is a Tertiary monzonite. The pitchblende is in a tension fracture (the Radium vein) of an inter-connecting vein system, at vertical depths between 875 and 1075 feet. Two pitchblende ore shoots are known; the largest one has a vertical dimension of 200 feet and a horizontal dimension of 70 feet. Within the ore shoots, pitchblende forms a discontinuous seam as much as 6 inches thick along the side of sulfide veins. The sulfides and the pitchblende are not intergrown. There is no apparent difference in mineralogy, other than the presence of pitchblende, between the pitchblende-bearing vein and the other veins in the system. The pitchblende is predominantly a soft, sooty variety, but the larger masses also contain a hard, botryoidal variety. The soft pitchblende is believed to have been deposited later and at a lower temperature than the hard pitchblende.

BRANCHING RATIO OF K^{40} RADIOACTIVE DECAY AND ITS APPLICATION TO GEOLOGICAL AGE DETERMINATIONS

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The ratio of K-capture to beta-emission for the naturally occurring radioactive isotope K^{40} has been determined by extracting, purifying and measuring the volume of argon extracted from four different samples of microcline. The amount of radiogenic A^{40} was ascertained by mass spectrometric analysis. The potassium contents of the samples of microcline were determined by chemical analysis. The argon was extracted by heating with metallic sodium in vacuum, and it was purified by treating the gas in a calcium furnace to constant volume. Isotopic analysis of argon thus obtained showed that the different samples contained from 93 to 100 per cent of radiogenic argon.

All the samples of microcline used in this investigation were from the Grenville area where ages ranging from about 850 to 1050 m.y. had been determined by uranium-lead methods. As the localities from which these samples were collected are not precisely dated, they have been assumed to be of the same age as Wilberforce (1030 m.y.). Using the total half life for K^{40} of 12.7×10^8 years as obtained by Sawyer and Wiedenbeck, the values of branching found were from 0.053 to 0.066. These are in good agreement with the low range of values obtained by counting experiments and also with those obtained by Aldrich and Nier. But if any of the sample had a younger age (for example 850 m.y.) then the values for the branching ratio would be raised and would still be within the range of other physical determinations.

Work upon more precisely dated material is being continued in the hope that the branching ratio can be more precisely fixed and a useful method of determining ages established.

α-EUCRYPTITE PROBLEM

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Massive granular α-eucryptite is reported from two new pegmatite localities, Center Strafford, New Hampshire, and the Harding mine, Dixon, New Mexico. The Center Strafford material is an intergrowth of α-eucryptite and albite, identical with that described by Brush and Dana (1880) from Branchville, Connecticut. The Harding material is white to pale gray with a waxy luster. Hardness $6\frac{1}{2}$. Specific gravity 2.64. Optically uniaxial positive (+) with $nO=1.573$ and $nE=1.583$. Analysis gave: Li_2O 8.36, CaO 0.19, Na_2O 0.62, K_2O 0.38, Al_2O_3 35.76, SiO_2 54.64, total 99.95; measured cell contents: $Li_{4.5}Al_{5.3}Si_{6.9}O_{24}$. The strongest x-ray powder lines: 3.35 (10), 3.93 (9), 1.37 (9), 2.74 (8).

Crystals of synthetic α-eucryptite, averaging 5 to 8 mm. in length and 0.5 to 1.0 mm. in width were made hydrothermally (by G. C. Kennedy). Hardness ~6. Hexagonal-R; hexagonal scalenohedral. Observed forms: $c\{0001\}$, $m\{10\bar{1}0\}$, $a\{11\bar{2}0\}$, $r\{10\bar{1}1\}$, $e\{01\bar{1}2\}$, $f(02\bar{2}1\}$, $v\{21\bar{3}1\}$. Unit cell dimensions: $a_0=13.54$ Å, $c_0=9.01$ ($a_0:c_0=1:0.6654$); space group: $D^5_{3d}-R\bar{3}m$. The optical and x-ray powder data are identical with those of the Harding material. Analysis gave: Li_2O 9.65, Al_2O_3 43.75, SiO_2 46.70, total 100.10; measured cell contents: $Li_{4.9}Al_{6.5}Si_{5.9}O_{24}$. Treating the composition as derivative from SiO_2 (or $Si_{12}O_{24}$), the general formula can be written $(Si_{12-x}, Al_x)(Li_{x-y}Al_{y/3})O_{24}$, where $x=5.1$ and $y=0.6$ in the Harding material, and $x=6.1$ and $y=1.2$ in the synthetic material. The mechanism of compositional variation is interesting in that there is a coupled interstitial entrance of 3Li+Al to provide valence compensation for Si=Al instead of alkali alone as in the tridymite-christiansenite and high-cristobalite—alpha-carnegieite series.

Both natural and synthetic α-eucryptite fluoresce begonia rose (Ridgway) in short-wave ultraviolet radiation. In the direct x-ray beam the natural material fluoresces pale shrimp pink; the synthetic, begonia rose. When irradiated, both become pale smoky.

PALERMOITE AND GOYAZITE, TWO STRONTIUM MINERALS FROM THE PALERMO MINE, NORTH GROTON, NEW HAMPSHIRE

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Palermoite, a new species having the composition $(Li, Na)_4SrAl_9(PO_4)_8(OH)_9$, and goyazite, $SrAl_3(PO_4)_5H_2O$, occur as late hydrothermal products in open cavities at the Palermo pegmatite, North Groton, New Hampshire.

Palermoite is orthorhombic, long prismatic in habit with {001}, {010}, {011}, {130} and {110}. The prism zone is vertically striated. Unit cell dimensions: $a_0=7.31$ Å, $b_0=15.79$, $c_0=11.53$ ($a_0:b_0:c_0=0.4630:1:0.7320$) with the space group D_{2h}^{25}-Imm. Cleavage {100} perfect, {001} fair. Fracture fibrous to subconchoidal. Brittle. Hardness 5½. Specific gravity 3.22 (meas.), 3.20 (calc. for two formula-units per cell with Li:Na=4.9:1). Colorless to white. Luster vitreous to subadamantine. Streak white. Fluoresces white in direct x-ray beam. Optically biaxial negative (−) with $n\text{X}=1.627$, $n\text{Y}=1.642$, $n\text{Z}=1.644$; $2\text{V}=\sim 20°$; $r<v$, moderate; $\text{X}=c$ (elongation), $\text{Y}=a$, $\text{Z}=b$. Analysis gave: Li_2O 3.70, Na_2O 1.56, K_2O 0.10, CaO 0.88, SrO 9.20, Al_2O_3 33.85, P_2O_5 44.64, H_2O 5.97, total 99.90. The strongest x-ray powder lines are 3.11 (10), 4.38 (9), 2.44 (8).

Small colorless to white rhombohedral crystals of goyazite, pseudocubic $\{10\bar{1}2\}$ terminated by {0001} and truncated by $\{20\bar{2}1\}$ and $\{11\bar{2}0\}$ occur in association with palermoite. Unit cell dimensions on Palermo crystal: $a_0=6.98$ Å, $c_0=16.54$ ($a_0:c_0=1:2.3696$). Specific gravity 3.15. Optically uniaxial positive (+) with $n\text{O}=1.640$ and $n\text{E}=1.651$, showing anomalous biaxial character with 2V up to 20°. The strongest x-ray powder lines: 2.97 (10), 2.20 (8), 1.89 (7).

The mineral association includes siderite, childrenite-eosphorite, green fibrous beraunite, crandallite-deltaite (?), whitlockite, brazilianite, tiny white hexagonal prisms of apatite, and small quartz crystals. Palermoite and goyazite appear in the sequence after siderite, but before apatite and quartz.

X-RAY STUDY OF RENIÉRITE, GERMANITE, AND COLUSITE

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X-ray powder photographs of reniérite, germanite, and colusite show them to be essentially isostructural. There are slight but consistent differences in the edge of the unit cells, and reniérite further shows some divergence from a true isometric pattern, accounting for the strong anisotropism observed in polished section. The average values of a_0, using both CuK_α and CoK_α filtered radiation, are as follows:

Colusite 10.62 ± Å, germanite and reniérite 10.57 ± Å. The value for germanite is twice the published figure (5.299 Å), and that for reniérite is new.

Intensities for all three correspond very closely, with the exception of one group of lines, where there is a consistent difference.

		Intensity	
Form	Colusite	Germanite	Reniérite
{002}	2	1	1
{012}	2	½	1
{112}	2	2	2

All lines can be satisfactorily indexed except in reniérite, where members of definitely resolved doublets at 1.33±Å, 1.08±Å, and .9390±Å cannot be so indexed, and suggest strongly that the mineral is pseudo-isometric.

MINERALOGY OF THE SERPENTINE GROUP

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An investigation of the stability, morphology and structure of a number of serpentine and serpentine-like minerals has resulted in a more critical evaluation of the fundamental properties of the end members chrysotile and antigorite, and in more information as to the nature of certain related minerals.

Chrysotile asbestos is less stable than antigorite as shown by treatment with HCl and analysis of the results by *x*-ray and differential thermal patterns, electron micrographs, and solubility data. Since nitrogen adsorption measurements reveal no significant difference in the surface area of the test samples, it is believed that the difference in solubility may be the result of greater strain in the outer layers of chrysotile tubes than is present in the flat antigorite plates. Hydrothermal treatment of finely ground chrysotile fibers produces a flaky substance resembling antigorite in its properties.

Using the effect of 1N HCl treatment upon the 001 *x*-ray reflection, a method was devised statistically which provides a means of evaluating the percentage of chrysotile and antigorite present in mixtures. With the aid of this method it has been shown that serpentines other than chrysotile and antigorite are usually mixtures of these two minerals.

A fibrous form of bementite has been found to be a manganese mineral analogous with chrysotile and ectropite may be a similar analog of antigorite.

Deweylite seems to consist of a crystalline serpentine phase plus an amorphous complex of alumina, silica, and water.

GEOPHYSICAL INVESTIGATIONS IN THE EMERGED AND SUBMERGED ATLANTIC COASTAL PLAIN. PART VII, THE CONTINENTAL SHELF, CONTINENTAL SLOPE, AND CONTINENTAL RISE SOUTH OF NOVA SCOTIA

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The results of a series of twenty-five refraction profiles made over the continental shelf south of Nova Scotia and extending down to the adjacent deep basin are reported. The shelf profiles are similar to those obtained elsewhere along the Atlantic Coastal plain, showing an unconsolidated sediment, a semiconsolidated sediment and a crystalline basement. The sediments gradually increase in thickness seaward up to 2.0 km (6500 ft.). On the continental shelf southeast of Halifax, Nova Scotia, a lens of higher velocity sediment is present below the semi-consolidated sediment with a maximum thickness of 2.7 km (9000 ft.). There is a downbuckle in the basement under the continental slope and a corresponding increase in the total thickness of sediments up to 5.0 km (16,000 ft.) on the continental rise. The sediments then thin further seaward with the younger sediments overlapping the older. On the profiles over the continental rise the intermediate layer of seismology was measured at an average depth of 10 km (33,000 ft.) and an average velocity of 6.9 km/sec (23,000 ft/sec), and on one profile the Mohorovicic discontinuity was measured at a depth of 16 km (52,000 ft.).

EOCENE MAGMA SERIES OF THE BEARPAW MOUNTAINS, MONTANA*

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In the Bearpaw Mountains region of north-central Montana, early Eocene deposition of terrestrial sediments, probably equivalent to the Wasatch formation, was halted by uplift due to irruption of magma that continued intermittently through the rest of Eocene time. Four major volcanic formations and a great variety of intrusive rocks are provisionally assigned to two magma series. One of these series is represented by igneous rocks related to a monzonitic petrographic suite and the other by igneous rocks related to a hyperalkalic shonkinitic petrographic suite. Metalliferous deposits and inclusions of pre-Cambrian basement rocks occur in rocks of the monzonitic series. Rocks of both series occur also in neighboring mountains and can be correlated with those in the Bearpaw Mountains. Evidence supports the thesis that the undersaturated shonkinitic magma, through assimilation of basement rocks, gave rise to the magma series that formed the rocks of the monzonitic suite.

NOTES ON THE DIFFERENTIAL LEACHING OF URANIUM, RADIUM, AND LEAD FROM PITCHBLENDE IN H_2SO_4 SOLUTIONS

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Two 5-pound samples from the "hot spot" of the pyritic dump of the Wood mine, a past producer of pitchblende near Central City, Colo., showed Ra/U ratios that were abnormally high but nearly constant at about 150 times the equilibrium value for both samples in spite of a seven fold difference in uranium contents. Analyses of oxidized but still black Katanga pitchblende before and after leaching in very dilute, dilute, and concentrated sulfuric acid solutions showed that: (1) UO_3 is preferentially leached with respect to UO_2, Ra, and Pb in all three solutions, (2) the resulting residual concentration of both radium and lead effected in the process is proportional to the total amount of uranium leached except in concentrated H_2SO_4, and (3) after leaching in concentrated H_2SO_4 the resulting residual concentration of radium relative to lead is much lower, as might be expected from data published by Lind, Underwood, and Whittemore in 1918 for the solubility of $RaSO_4$. Under similar leaching conditions unaltered pitchblende from Great Bear Lake lost only 1/10 to 1/15 as much uranium as the UO_3-rich Kantanga ore. Both laboratory and field results point to the same conclusion: in an oxidizing, highly acid environment uranium is rapidly leached and both radium and lead tend to be fixed approximately proportionally in the process. These results help to explain (1) why UO_3-rich uranium minerals tend to give *maximal* Pb/U ages and (2) why the search for high-grade uranium ore, in and around abandoned sulfide mines known to have produced pitchblende in the past has been consistently disappointing.

ABUNDANCE OF Ba, Sr, Sc AND Zr IN CHONDRITES AND ULTRAMAFIC ROCKS

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Twenty one chondritic meteorites and a number of mineralogically related ultramafic rocks have been analyzed by the optical spectrograph for some elements believed to be restricted largely to the silicate phase of meteorites. The concentration of each of these elements remains remarkably uniform in chondrites, lending evidence to the theory that meteorites had a common origin in both space and time. Analytical results and their implications in the Shattered Planet Hypothesis of the origin of meteorites are discussed.

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* Publication authorized by the Director, U. S. Geological Survey.

In chondrites the following abundance values were obtained: Ba 0.0008%, Sr 0.0007%, Sc 0.0007% and Zr 0.005%. The validity and accuracy of these results is discussed in some detail.

Some preliminary observations have been made on the abundance of lithium in chondrites and ultramafic rocks. This element shows a strikingly large abundance difference in these two materials and its concentration in chondrites is more than twenty times greater than that in ultramafics. Ba, Sr, Sc and Zr do not show such extremely large abundance differences.

SEISMIC REFRACTION MEASUREMENTS ON THE GRAND BANKS AND ADJACENT SHELVES

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A reconnaissance refraction survey of the Grand Banks, St. Pierre Banks, and Banquereau was conducted in the first two weeks of September 1951. Three partially reversed profiles on the Grand Banks indicate sedimentary layers ranging in total thickness from 2000–4500 feet, underlain by crystalline rocks with velocities of 16,000–18,000 ft./sec. On St. Pierre Bank a single partially reversed profile showed 8000 feet of sediment underlain by crystalline rocks with a velocity of 17,500 ft/sec. Similar results were obtained on Banquereau where 9000 feet of sediments were found overlying crystalline rock with a velocity of 15,500 ft/sec. A partially reversed profile in the channel between St. Pierre Bank and Banquereau indicates 15,000 feet of sediment overlying crystalline rock with a velocity of 20,500 ft/sec. These results suggest that the channel is structural in origin.

WIDESPREAD OCCURRENCE AND CHARACTER OF PITCHBLENDE IN THE TRIASSIC AND JURASSIC SEDIMENTS OF THE COLORADO PLATEAU

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During the last three years numerous uranium deposits in the sedimentary rocks of the Colorado Plateau have been shown to contain pitchblende. Although this mineral is not restricted to any one formation, the majority of its occurrences are in the lower Chinle and Shinarump formations of Triassic age. There are two principal modes of occurrence of pitchblende, one with sulfides of copper, the other in asphaltic bodies. In both types, the association with fossil plants or the occurrence in beds rich in organic debris is generally the rule. Fossil plants replaced by pitchblende and sulfides show a remarkable preservation of the organic structures. Pitchblende generally replaces cell walls, while the copper sulfides fill the cells, as is shown particularly well at White Canyon, Utah. Under these conditions the hardness and reflectivity of the pitchblende may differ considerably from that of hydrothermal vein pitchblende. At Seven Mile Canyon, Utah, pitchblende is found as rims on quartz and limestone pebbles as well as replacing wood. The ores of Temple Mountain, Utah, contain pitchblende as minute grains, one to ten microns in diameter, arranged in ribbon or ellipsoid-like aggregates in asphaltite. It is likely that not all of the uranium occurs as pitchblende, a portion being combined with the organic matter. The paragenesis of the minerals is complex and somewhat obscure, but a formation of the pitchblende contemporaneous with, or shortly followed by, copper sulfides is suggested. In the asphaltite deposits a polymerization by alpha particles of light hydrocarbons may be involved.

STUDIES IN THE SYSTEM $MgO-Al_2O_3-SiO_2-H_2O$

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Continuing earlier hydrothermal studies of the systems $MgO-Al_2O_3-H_2O$ and $Al_2O_3-SiO_2-H_2O$, the influence of silica and magnesia on these respective ternary systems was studied. Magnesian serpentine is shown to admit about 10 mol. per cent Al_2O_3 into solid solution, thereby raising its temperature of stability at least 15° and changing its habit from tubular to platy. The preparation of various serpentines including nickel serpentine (slightly fibrous) and magnesia-germania serpentine (good hexagonal plates) lend support to the concept that size of ions controls the morphology. Pure aluminous montmorillonite admits magnesia into its structure, giving rise to a large primary phase volume of quaternary solid solutions. The decomposition temperature of magnesian montmorillonite is greater than that of the pure aluminian end member, attaining a maximum temperature of about 500° C. An "X-phase" is also encountered.

Phase equilibria in the high-alumina corner of the system indicate stability of the corundum-cordierite join above 500° C. and 5,000 psi. Apparent disagreement with earlier work in other parts of the system and the effect of the nature of the starting material on the attainment of equilibrium are discussed.

MINERALOGY OF THE RESIDUAL KAOLINS OF THE SOUTHERN APPALACHIANS

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An investigation was made of the residual kaolins of the Southern Appalachian region to evaluate conditions giving rise to halloysite and kaolinite.

A new differential thermal analysis technique was developed to determine quantitatively the relative amounts of endellite versus kaolinite and/or halloysite in a sample. This was used along with other analytical methods to make quantitative mineralogical analyses of the residual clays.

The mineralogical analyses and thin section studies formed the basis for determining the sequence of weathering in the granites and pegmatitics. Halloysite is formed only from the weathering of feldspars, and under conditions of intense leaching is formed from both the plagioclase and potash feldspars. The primary muscovite always alters to vermicular kaolinite. Where leaching is less intense, vermicular kaolinite is formed from the potash feldspars through the intermediate stage of secondary mica of the muscovite type and halloysite is formed from the plagioclase feldspars. Where acid leaching conditions obtain but are least intense, vermicular kaolinite is formed from the weathering of feldspars through the intermediate stage of secondary mica with minor amounts of halloysite forming directly from feldspar. Once secondary mica is formed, vermicular kaolinite is the next weathering product.

PROBLEM OF TRACE-ELEMENT DISTRIBUTION IN MAGMATIC MINERALS

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Goldschmidt and his associates showed that the distribution of trace-elements in magmatic minerals depends on the camouflage principle, whereby a minor element replaces a major element of similar size. It was also claimed that where there is a difference in size the larger element (cation) is concentrated in later fractions of the mineral. Evidence is presented to show that this latter statement is not a general rule and has no theoretical or experimental basis. Camouflage may also occur where there is a valency difference, but again it cannot be predicted from valency and size alone, whether the minor element is concentrated or dispersed during magmatic crystallization.

Theory shows that where a minor element, e.g., Tl^{+1} replaces K^{+1} in the potash feldspars (continuous solid solutions assumed), then the M.P. of the hypothetical feldspar $TlAlSi_3O_8$ determines whether thallium is enriched or impoverished during crystallization. Melting-points have no direct relation to ionic size. Collected data on the M.P.s of end-members in binary solid solution series confirm that there is in fact no regular relation between M.P. and ionic size or charge of the cation. If conditions are assumed to be more complex than the simple binary solid solutions, ionic size and charge matter even less. Wickman's activation energies are of little assistance, because they are closely related to lattice-energies, which do not determine melting-points.

The camouflage, capture and admission principles were established on the basis of spectrographic data on minor element abundances. Review of the more recent literature shows again that these principles are incapable of general application. Where they seem to apply (e.g., Li and Mg, Tl and K) this must be regarded as coincidental. In conclusion, it is at present impossible to predict the behaviour of a minor element during magmatic evolution from ionic properties alone, beyond indicating generally where it will be camouflaged.

LEAD-URANIUM AGES OF SOME URANINITES FROM TRIASSIC AND JURASSIC SEDIMENTARY ROCKS OF THE COLORADO PLATEAU

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In the course of a study on the origin of the uranium deposits in the Triassic and Jurassic sedimentary rocks of the Colorado Plateau, 13 samples of uraninite from nine deposits were collected for Pb^{206}/U^{238} and Pb^{207}/U^{235} age determinations. These uraninites are believed to be the most reliable of more than 100 samples from the Plateau on which age determinations have been made. The 13 uraninites have a mean Pb^{206}/U^{238} age of approximately 70 million years when corrected only for common lead.

Chemical and mass spectrometric errors change the mean Pb^{206}/U^{238} age by approximately ±4 million years. Uncertainties resulting from the correction for common lead and from the possible presence of old radiogenic lead in the ores will decrease the mean Pb^{206}/U^{238} age by approximately 3 million years. Corrections for the selective loss of uranium will decrease the mean Pb^{206}/U^{238} age, whereas selective loss of daughter products will increase the mean Pb^{206}/U^{238} age by approximately 10 million years.

Holmes gives 130 and 160 million years as the close of the Jurassic and Triassic periods, respectively. If the ages calculated for the samples of uraninite are close to the true ages of the ores, then the uranium was introduced into the sedimentary rocks in late Mesozoic or early Tertiary time (60 to 70 million years ago). This conclusion differs markedly from conclusions based on other evidence that the uranium ores were formed in the Triassic and Jurassic sedimentary rocks of the Colorado Plateau during or soon after the deposition of the rocks.

PORPHYRY INTRUSIONS IN THE BEARTOOTH RANGE, NEAR RED LODGE, MONTANA

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Porphyry intrusions are exposed for a vertical distance of about 4,000 feet, at approximate elevations between 6,000 and 10,000 feet along the Red Lodge–Cooke City Highway into Yellowstone Park. As it is unusual to find such fresh and deeply cut exposures, the the present study has been begun to see if mineral or textural change occurs with depth. This may aid in understanding the method of intrusion or the relationship of the porphyry to the cover rock.

The porphyries form concordant sheets or sills, ranging from 50 to several hundred feet thick. Their attitude varies from almost horizontal to about 30° dip. Jointing is pronounced. Faulting occurred within the sheets, and some contacts with pre-Cambrian rocks are faults marked by gouge and mylonite zones.

The rock composition is quartz monzonite. The fine-grained ground mass is either brown, gray or light gray with phenocrysts of feldspar (abundant), quartz and hornblende. In the limited section studied along the highway and on Line Creek Plateau, the porphyries are in contact with pre-Cambrian granites, gneisses and amphibolites. The field survey shows no mineral or textural change in the rock with depth. A petrographic examination is in progress to see whether a variation occurs. The work also will be extended to examine porphyry bodies which have intruded sediments in nearby areas.

The porphyries examined are younger than the pre-Cambrian rocks. Similar porphyries in associated areas, which have intruded sediments, have been dated by others as late Cretaceous or very early Tertiary in age.

INTRUSION BRECCIAS OF MAJUBA HILL, NEVADA

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Majuba Hill, in the Antelope Range, Pershing County, Nevada, is a complex, roughly elliptical plug that has a maximum diameter of 5,000 feet. Shale, limestone, and sandstone of Triassic (?) age was first invaded by Tertiary (?) dikes and sills of andesite, dacite, latite, and diorite, and later by the plug of rhyolite prophyry, quartz porphyry, and porphyritic rhyolite.

Dikes and irregular bodies of breccias cut all rocks within the plug except the porphyritic rhyolite and cut the metamorphosed shales along the borders. More than 20 lithologic types of intrusion breccias are recognized. Rhyolite porphyry, quartz porphyry, or a mixture of quartz and tourmaline commonly forms the groundmass of the breccias; and copper, tin, and uranium minerals occur in the intrusion breccias that contain a matrix of quartz and tourmaline.

The intrusion breccias are believed to have been formed by: (1) incorporation and incomplete assimilation along the edges of rhyolite porphyry and quartz porphyry dikes, (2) autobrecciation of rhyolite porphyry and quartz porphyry dikes and masses as they were injected into the plug, and (3) minor explosive action resulting from the rapid release of gases emanating from the magma.

MASS SPECTROMETRIC DETERMINATION OF THORIUM

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Thorium possessing a Th^{230} to Th^{232} ratio of four has been isolated in milligram amounts from a Colorado Plateau uranium ore. This carrier has been used in conjunction with a mass spectrometer equipped with an electron multiplier which permitted isotopic analysis of amounts of thorium as low as one microgram. Under these conditions it was possible to determine 0.01 microgram of thorium with an accuracy of $\pm 10\%$ and amounts of 0.1 microgram or more to $\pm 1\%$.

The method has been applied to zircon in a granite from the Haliburton Area of the Canadian Shield. Three analyses of separate powder fractions gave results of 2012, 2033 and 2058 parts per million of thorium with an analytical error of $\pm 1\%$ for each determination.

On the basis of previously determined uranium and lead data the age of the mineral is believed to be 1050 ± 20 million years. The newly calculated thorium-lead age is 420 ± 10 million years. A sufficient number of uranium, thorium and lead analyses have been made to establish that the sampling was representative. Since the zircon has metamict and unaltered zones the data may be explained by an accumulation of thorium relative to uranium in the metamict phase, which could allow Pb^{208} to be lost more easily than Pb^{206} or Pb^{207}. Further investigation of the problem is under way.

THERMAL TRANSFORMATIONS IN SOME LAYER SILICATE MINERALS

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Single crystals of muscovite, biotite, and chlorite and sedimented flakes of illite were subjected to progressive, stepwise heat treatment by two methods: (1) conventional air quench procedure, and (2) high temperature method with furnace mounted on an *x*-ray spectrometer. Comparisons of the *x*-ray data indicate marked differences in the high temperature region (800–1100° C.) depending on variations in experimental procedure, particle size, and perhaps other factors. The high temperature furnace procedure was found to be the more rapid and accurate method.

Visual images of the structure (electron density along *c*-axis) were calculated by Fourier syntheses. The close relationship between the specific layer structure, the amount of dehydration, the gradation (or lack of it) to an anhydride, and the early appearance of new phases were observed.

Precise evaluation of thermal transformations requires careful consideration of the time factor with respect to particle size and degree of crystallinity, approach to equilibrium, and decomposition temperature.

SIGNIFICANCE OF THE VARIATION IN NORMATIVE vs. MODAL FELDSPAR OF GRANITES

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Granites having identical normative amounts of "plagioclase" and "orthoclase" may contain equivalent quantities of these constituents or they may consist of "one" feldspar (perthite) and quartz. Modal analyses of a series of specimens collected along the contacts of tertiary granites from the Island of Skye, Scotland, show a variation in modal plagioclase from 0 to 20 per cent. Chemical analyses of the granites indicate that the normative feldspar composition is approximately 30 per cent "plagioclase" and 30 per cent "orthoclase." Such variations in modal feldspar suggest, (1) that the granites crystallized at different temperatures, or, (2) that the plagioclase unmixed from the potash-bearing feldspar subsequent to primary crystallization. Evidence will be presented for and against each of these possibilities.

PETROLOGY OF THE JUDITH MOUNTAINS, MONTANA

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The Judith Mountains are a group of Tertiary porphyry stocks and related intrusives that have domed and cut a thick series of sedimentary rocks ranging in age from Cambrian to Upper Cretaceous. Quartz monzonite intrusions mark the beginning of intrusive activity and form the major stocks in all parts of the mountains. Later igneous rocks represent two lines of descent: (1) more siliceous types exemplified by rhyolite, (2) rocks relatively poor in silica and high in the alkalies, best exemplified by tinguaite. The youngest rock in the area, the alkali granite porphyry of Judith Peak, is rich in both silica and the alkalies, and is believed to represent a union of the two separate lineages.

The area is bisected by a strong but irregular east-west fault that divides the mountains into two structural units, each with a distinct assemblage of igneous rocks. The southern unit is characterized by calc-alkaline rocks containing perhaps 50 per cent of highly siliceous rocks, i.e., the rhyolites. The northern unit is characterized by both calc-alkaline and alkaline rocks. Only a small percentage, probably less than 1 per cent, of the calc-alkaline rocks is rich in silica.

Pre-intrusive structural and erosional events have influenced the mechanism of emplacement of the initial quartz monzonite intrusions. Detailed mapping shows large-scale magmatic sloping of the Madison limestone north of the strong east-west fault but not south of it. This suggests that limestone syntexis has desilicated the magma at depth to yield tinguaites, according to Daly's hypothesis. The alkali granite is believed to represent a late siliceous differentiate of the calc-alkaline series, contaminated by alkaline fluids generated by resurgent carbon dioxide released from the assimilated limestone.

MINERALOGIC STUDIES ON THE COLORADO PLATEAU

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The primary studies on the mineralogy of the Colorado Plateau, such as the establishment of standards for *x*-ray identification of uranium and associated minerals, techniques for separating and studying radioactive minerals, and acquaintance with Colorado Plateau minerals and mineral assemblages, have reached the stage where the mineralogic study of the Plateau uranium ores is progressing rapidly. About 20 uranium minerals and several uranium-bearing minerals or organic materials have been identified, as well as many associated minerals. The principal uranium minerals are uraninite and pitch-blende, hydrous oxides, carbonates, sulfates, phosphates, arsenates, vanadates, and silicates. The chief metals associated with the uranium are vanadium and/or copper. The minor metals are lead, zinc, cobalt, nickel, manganese, silver, molybdenum, gold, arsenic, and selenium. Hummerite and montroseite, two new vanadium minerals, have been described. A possible new uranium mineral and several other possible new vanadium minerals are being investigated.

Recent work suggests that carnotite ore is restricted to near-surface deposits in the zone of oxidation. Black uranium-vanadium ore occurs in scattered surface deposits and in the few deep deposits so far known in western Colorado. It is less oxidized and contains more sulfides than the carnotite ore. Present study of several new minerals from the black ore is concerned with the paragenesis of the uranium-vanadium ores. The minerals of areas containing copper, of which the White Canyon district, San Juan County, Utah, has been most thoroughly studied, are uraninite, copper and iron sulfides (with minor amounts of galena), sphalerite, cobalt and nickel minerals, and many secondary hydrous oxides, carbonates, and sulfates of uranium and other metals. Altered wall rock in the White Canyon district contains jarosite, alunite, kaolinite, allophane, and gibbsite.

FLUID INCLUSIONS IN CERTAIN MINERALS FROM THE ZONED PEGMATITES OF THE BLACK HILLS, SOUTH DAKOTA, AND THEIR SIGNIFICANCE

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A study of fluid inclusions in minerals from zoned pegmatites was undertaken as a check on the theory of zonal development through fractional crystallization in a restricted system. Beryl, spodumene and garnet gave results that showed general agreement; the most extensive results were obtained from beryl.

Temperature determinations were made by heating polished mineral sections under the microscope and recording the temperature of disappearance of the vapor phases of the fluid inclusions.

Beryl from outer zones of the Highland Lode and Bob Ingersoll No. 1 pegmatites yielded higher temperatures than beryl from inner zones. Results from the Peerless and Bob Ingersoll No. 2 pegmatites were inconclusive or contradictory. No other pegmatites yielded enough data to allow establishment of any temperature trends.

Temperatures observed in beryl ranged from 216° to 515° C. Results from individual crystals showed a range of 9° to nearly 200° C., and individual zones gave ranges of 65° to 230° C.

Data were too meager to permit presentation of conclusions, but results give tentative support to the hypothesis of fractional crystallization. Estimates of pressures prevalent at the time of crystallization imply temperatures of crystallization in excess of 800° C. for a few of the early-forming beryl crystals.

PRACTICAL NOTATION AND ARRANGEMENT OF THE CRYSTAL CLASSES

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The notations and arrangements of the 32 crystal classes adopted in current reference works can be criticised as inadequate in practical respects. They are not easy to remember, nor convenient to write and print, nor yet suited for consideration of macroscopic physical properties. We show a notation for the classes, based on modified international symbols for 13 kinds of macroscopic symmetry axes, and put into a 4 by 12 tabular arrangement, which we have found to be more practical in several respects. Some physical properties of interest in crystallography include for both isotropic and monocrystalline solids the following:

Voigt—Bhagavantam	Jahn	Constants	Examples
	Centric Phenomena		
vector-vector	$[V^2]$	16	Dielectric susceptibility, crystal optics, etc.
scalar-tensor	$[V^2]$	1–6	Thermal expansion.
tensor-tensor	$[[V^2]^2]$	2–21	Elasticity.
tensor-tensor	$[[V^2]^2]$	2–36	Photoelasticity.
	Acentric Phenomena		
scalar-vector	V	1–3	Hydrostatic piezoelectricity, etc.
pseudoscalar-tensor	$E[V^2]$	1–6	Optical rotation.
vector-tensor	$V[V^2]$	1–18	Piezoelectricity, electro-optics.

THREE-DIMENSIONAL PICTURE OF STEAMBOAT SPRINGS, NEVADA

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Carson City, Nevada

Diamond drilling by the Geological Survey has added immeasurably to an understanding of the structure, mineral deposition, rock alteration, temperature relations, and migration of water at Steamboat Springs.

The thermal history is long and involved, perhaps extending throughout Quaternary time. The dominant structure appears to be a west-dipping thrust fault, with springs emerging from east-dipping normal faults in the hanging wall of the thrust.

Six drill holes 400 to 686 feet deep indicate a maximum or near-maximum temperature for each hole at depths of 200 to 360 feet below water level; maximum temperatures are 318 to 342° F. At greater depths temperatures level off, or decrease slightly below main structures serving as channelways. The most satisfactory interpretation of these relations is: At considerably greater depths, a volcanic component (probably steam with dissolved substances) and meteoric water mix with proportions and temperatures such that a single liquid phase results, with a temperature probably little above 342° F. This water rises with minor loss of temperature by conduction, except on margins of the thermal area, until it is within several hundred feet of the surface. With further rise and decrease of pressure, a vapor phase forms. The high near-surface geothermal gradient is largely due to boiling. Dissolved gases tend to lower the boiling range below that of pure water, whereas dissolved solids have an opposite effect.

Near the surface some meteoric water mixes with thermal water on margins of the area. Hydrodynamic head and density, which is a function of temperature, are controlling factors.

SUTTER BUTTES RESTUDIED

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University of California, Berkeley, California

Since the work of Williams on the Marysville Buttes (1929—now officially called Sutter Buttes) evidence has been found modifying some of the ideas then expressed concerning their evolution.

It is no longer believed that they are a denuded laccolith, result of a single andesite intrusion, but a complex of andesite intrusions of at least three different ages which form a stock-like core 4 miles in diameter. Rhyolite plugs surrounding this core are not the last intrusive phase, though they are later than some of the andesite intrusions. Rhyolite tuff, however, found at the base of associated volcanic sediments surrounding the central core, marked the initial explosive phase and was erupted before appreciable deformation of pre-volcanic sedimentary rocks had occurred. The buttes themselves are andesite plugs of later age than the rhyolite plugs and probably later than the central vent-tuff, formerly thought to have been the final phase of activity. They protruded through earlier intrusions and sedimentary cover to form great spines from which originated the youngest, very coarse lahars. Andesite block-lava flows, originating at vents in the peripheral sedimentary ring, closed the volcanic epoch.

The slow accretion of the central core caused progressive deformation of pre-volcanic and volcanic sedimentary beds, resulting in an unconformity which transgresses all but the final deposits of volcanic debris.

PSEUDOSYMMETRY AND ATOMIC PATTERNS OF CRYSTALS

DOROTHY WRINCH
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One set of cases in which it can be demonstrated that the pseudosymmetry of a crystal of known composition provides a pointer to the general nature of the atomic pattern is exemplified by orthorhombic staurolite $H_2FeAl_4Si_2O_{12}$, for which $a:b:c=0.4734:1:0.3414 \sim\sqrt{2}:3:1=2:3\sqrt{2}:\sqrt{2}$. Here the companion cell (with vertices actually on a cubic lattice) has displacements z[002], s[330], s′[$\bar{1}$10] and volume 12, all in the scale of some metric k in the range $a/2$, $b/3\sqrt{2}$, $c/\sqrt{2}$. Another set is exemplified by orthorhombic chrysoberyl $BeAl_2O_4$ for which $a:b:c=0.470:1:0.580\sim\sqrt{2}:3:\sqrt{3}=\frac{2}{3}\sqrt{3}:\sqrt{6}:\sqrt{2}$, with the corresponding displacements $\frac{2}{3}$t[111], r[11$\bar{2}$], s[$\bar{1}$10] and volume 4, in the scale of a metric k in the range $\frac{1}{2}a\sqrt{3}$, $b/\sqrt{6}$, $c/\sqrt{2}$. Two interesting conclusions follow. (1) For staurolite, $12k^3$ provides $48=12\times4$ close-packed sites, suggesting that $Z=4$: for chrysoberyl, $4k^3$ provides $16=4\times4$ such sites, suggesting that here too $Z=4$. (2) That the close-packed sites for staurolite are cubic (i.e. fcc) is shown by the z, s, s′ displacements of the companion cell, since the required symmetry elements for this orthorhombic crystal associated with each of these directions (2, m, or 2/m) lie within the cubic system. On the other hand, such symmetry elements associated with r and t displacements do not lie within the cubic system. We infer from the r, s, t nature of the displacements for chrysoberyl that the close-packed sites in this crystal are of double cubic, not of cubic type.

DIOPSIDE-WATER SYSTEM

HATTEN S. YODER, JR.
Geophysical Laboratory, Washington, D.C.

The diopside-water system has been investigated up to 6500 bars by means of sealed platinum tubes held in an internally heated, high-pressure vessel. The melting point of diopside is lowered from 1391°C. at atmospheric pressure to 1289° C. at the maximum pressure of water vapor applied. This lowering is considerably less than the measured lowering(∼400°) of the melting point of albite under the same vapor pressure of water.

It was not feasible to quench the melt to a glass under the conditions of most of the experiments, hence a quantitative measure of the water content of the liquid could not be obtained. However, the preliminary experiments support the view that a basaltic magma contains much less water than a granitic magma.

LIST OF FORMER OFFICERS AND MEETING PLACES

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1927 Arthur L. Parsons, William F. Foshag, William A. Tarr, Alexander N. Winchell.
1928 William F. Foshag, William A. Tarr, Alexander N. Winchell, Ellis Thomson.
1929 William A. Tarr, Alexander N. Winchell, Ellis Thomson, Clarence S. Ross.
1930 Alexander N. Winchell, Ellis Thomson, Clarence S. Ross, Paul F. Kerr.
1931 Ellis Thomson, Clarence S. Ross, Paul F. Kerr, William S. Bayley.
1932 Clarence S. Ross, Paul F. Kerr, William S. Bayley, William J. McCaughey.
1933 Paul F. Kerr, William S. Bayley, William J. McCaughey, Kenneth K. Landes.
1934 William S. Bayley, William J. McCaughey, Kenneth K. Landes, E. P. Henderson.
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1938 J. F. Schairer, Arthur F. Buddington, Arthur P. Honess, R. C. Emmons.
1939 Arthur F. Buddington, Arthur P. Honess, R. C. Emmons, Carl Tolman.
1940 Arthur P. Honess, R. C. Emmons, Carl Tolman, D. Jerome Fisher.
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1948 R. E. Grim, Joseph Murdoch, H. H. Hess, Clifford Frondel.
1949 Joseph Murdoch, H. H. Hess, Clifford Frondel, Lewis S. Ramsdell.
1950 H. H. Hess, Clifford Frondel, Lewis S. Ramsdell, E. F. Osborn.
1951 Clifford Frondel, Lewis S. Ramsdell, E. F. Osborn, George T. Faust.
1952 Lewis S. Ramsdell, E. F. Osborn, George T. Faust, Victor T. Allen.

Annual Meeting Places

1920 Chicago, Illinois
1921 Amherst, Massachusetts
1922 Ann Arbor, Michigan
1923 Washington, D. C.
1924 Ithaca, New York
1925 New Haven, Connecticut
1926 Madison, Wisconsin
1927 Cleveland, Ohio
1928 New York, N. Y.
1929 Washington, D. C.
1930 Toronto, Canada
1931 Tulsa, Oklahoma
1932 Cambridge, Massachusetts
1933 Chicago, Illinois
1934 Rochester, New York
1935 New York, N. Y.
1936 Cincinnati, Ohio
1937 Washington, D. C.
1938 New York, N. Y.
1939 Minneapolis, Minnesota
1940 Austin, Texas
1941 Boston, Massachusetts
1942 No meeting held
1943 No meeting held
1944 No meeting held
1945 Pittsburgh, Pennsylvania

1946 Chicago, Illinois
1947 Ottawa, Canada
1948 New York, N. Y.
1949 El Paso, Texas
1950 Washington, D. C.
1951 Detroit, Michigan
1952 Boston, Mass.

Recipients of the Roebling Medal

Charles Palache, December 1937
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MINERALOGICAL SOCIETY (London)

The anniversary meeting of the Society was held on Thursday, November 6th, 1952, at 5 P.M., in the apartments of the Geology Society of London, Burlington House, Piccadilly, W.1. (by kind permission).

The following papers were read:—

(1) IMPERFECTIONS IN THE CRYSTAL STRUCTURE OF KALIOPHILITE

By Dr. A. M. B. Douglas.

Kaliophilite with anomalous optical properties has been recorded by Professor C. E. Tilley. Single-crystal *x*-ray investigations show that the basic crystal structure is identical with that of normal kaliophilite, but anomalous effects are observed which can be explained as due to imperfections in the structure. These could also account for the anomalous optical properties, and the imperfect structure is closely related to the orthorhombic form of $KAlSiO_4$.

(2) MORPHOLOGY OF BARYTE

By Mr. A. F. Seager.

A morphological analysis of barytes has been made in the manner of Donnay and Harker, using a modification of their method of estimating the relative importance of forms. An attempt has been made to predict the space group. A comparison of the observed and theoretical morphological aspects is given.

(3) GYROLITE

By Dr. A. L. Mackay and Dr. H. F. W. Taylor.

X-ray single crystal investigations on gyrolite ($2CaO \cdot 3SiO_2 \cdot 2H_2O$) show that it is hexagonal, with $a = 9.72 \pm 0.01$ Å, $c = 132.8 \pm 0.3$ Å, space group $C6_1(C6_5)$, $N = 48$. The dehydration curve shows that three-quarters of the water is lost much more easily than the remainder, and it is therefore possible that the formula of gyrolite should be written $Ca_4(Si_6O_{15})(OH)_2 \cdot 3H_2O$. The unit cell probably contains six chemically discrete layers, stacked parallel to the *c*-axis, and of thickness 22.13 Å. The structure of these layers is still uncertain, the *a*-axis being apparently too long to permit of a structure containing $(Si_2O_5)n$ sheets of the mica type. The dehydration of single crystals of gyrolite has been studied by *x*-ray methods. Loss of the first three-quarters of the water, at 450°, produces only minor changes in the structure, but loss of the remainder brings about conversion to α-wollastonite, even at temperatures considerably below that (1125°) at which it becomes the stable form. The changes from gyrolite to α-wollastonite, and also from the latter to the β-form on annealing at 1050°, take place in a completely ordered way, and it seems probable that a close structural resemblance between gyrolite and α-wollastonite must exist.

(4) AN UNUSUAL SYNTHESIS OF ACANTHITE CRYSTALS

By Dr. F. A. Bannister.

Description of an accidental growth of large transparent blades of acanthite by the diffusion of sulphur contained in plasticine to the silver electrodes of apparatus. The apparatus had been kept in closed metal vessels under normal conditions of temperature and pressure for ten years.

(5) A RE-EXAMINATION OF CHURCHITE

By Dr. G. F. Claringbull and Dr. M. H. Hey.

A new analysis of churchite shows that it is a hydrous yttrium phosphate with only minor proportions of cerium. It has been commonly regarded as essentially a cerium phosphate, but this is not supported by Church's original analysis (1865) which leaves the exact nature of the rare earths present in doubt. Weinschenkite first described in 1923 is identical with churchite. The name churchite has priority.

(6) Ferrous Chamosite and Ferric Chamosite

By Dr. G. W. Brindley and Mr. R. F. Youell.

A substantially pure ferrous chamosite has been obtained, which, on heating in air at 400° C., is converted to a wholly ferric form. The oxidation is accompanied by a partial dehydration of the structure which is analysed chemically and by *x*-rays. The external hydroxide sheet of the kaolin-type layer structure of chamosite is converted to a sheet of oxygen atoms. The internal hydroxide radicals remain in the lattice up to about 450°–500° C. The reduction of ferric to ferrous chamosite has been effected by heating at 460° C. in an appropriate H_2/H_2O mixture in absence of air. The oxidation process is accompanied by characteristic colour changes from green to brown. At higher temperatures up to 1100° C., ferrous aluminate, spinel, hematite and cristobalite are recorded depending on the conditions of heating.

The following papers were taken as read:—

(1) The Space-Group and Cell-Dimensions of Rankinite

By Mrs. K. M. Moody. (Communicated by Dr. W. H. Taylor.)

Oscillation, Weissenberg, and Laue *x*-ray photographs gave for rankinite from Scawt Hill, Co. Antrim: $a=10.55$, $b=8.88$, $c=7.85$ Å, β $120.1° \pm 0.5°$; sp. gr. 2.96, $4[Ca_3Si_2O_7]$.

(2) An Occurrence of Harmotome in N.W. Ross-shire

By Dr. C. D. Waterston. (Communicated by Dr. W. Campbell Smith.)

Harmotome crystals (average 3 mm. long) coat the vein-stuff in cracks and cavities of a fault breccia on the coast road between Lochinver and Ullapool, $\frac{1}{4}$ mile S.W. of the southern margin of Loch an Arbhair. The habit and properties of the crystals are described.

(3) The Ferriferous Orthoclases of Madagascar

By Dr. D. S. Coombs.

New chemical and optical data are presented for seven crystals of yellow ferriferous orthoclase from Madagascar. These are considered in conjunction with data taken from the literature, in an investigation of the seemingly anomalous determination of the optic axial plane parallel to (010), made by Kôzu in 1916 on a crystal from this locality. It is found that the entry of iron into the orthoclase structure tends to lower the optic axial angle at a rate of about 5° for 1% iron-orthoclase (0.25% Fe_2O_3 approximately). For iron contents beyond about 7.5% iron-orthoclase the optic axial plane is normally parallel to (010). A significant proportion of crystals have optic axial angles which are larger (in the plane perpendicular to (010)) than would be expected for this simple relationship, possibly indicating transitions to adularious forms. One crystal has been found with a small 2V parallel to (010) instead of perpendicular to (010) as expected from the suggested relationship.

(4) An *X*-Ray Method for Studying Orientation of Micaceous Minerals in Shales, Clays and Similar Materials

By Dr. G. W. Brindley.

The method is based on the measurement of the *x*-ray intensities of basal, 001, reflections round a Debye-Scherrer ring using a flat specimen set at an angle to the incident *x*-ray beam equal to the Bragg angle for each reflection considered. The main consideration is the correction for absorption in the specimen, which is discussed fully. The method is illustrated by *x*-ray diagrams of shales and artificially orientated clay specimens.

(*Titles and abstracts kindly submitted by G. F. Claringbull, General Secretary.*)

NOTES AND NEWS

PRECESSION ORIENTATION PHOTOS

D. Jerome Fisher, *University of Chicago, Chicago 37, Illinois.*

Precession orientation photos were discussed in a recent number of this journal in the first part of an article by the writer.[1] In what follows it is assumed that the reader is familiar with the material on pages 1037 to 1044 of this paper.

Once a correctly-oriented double exposure picture with dial values 180° apart (see page 1043) has been obtained with a given instrument, the relation of O' (the center of precession for the film; see page 1044) to the direct beam spot s' (see page 1039) is readily established. Thereafter on any new picture, O' can be pricked through the film after it has been in the fixer a minute or so, if it is placed above the polar net light box in such fashion that the Laue streaks lie along radii of the net, when s' will then be at the center of the net. Of course if there is no streak parallel the dial axis, the azimuth of the film under these conditions can be controlled by the cassette dots.

Actually the center c of the circle of precession should first be pricked on the wet film before establishing O' as outlined in the preceding paragraph. Then when O' is located, one can read from the polar net the angle A between the vector $O'c$ and the dial axis (cf. Fig. 40 *b*).[2] If the length (in mm.) of $O'c$ is measured, the orientation error ϵ can be obtained from Fig. 39, since $O'c$ is the same as (δ-r) of this graph.[3] If M' marks the point where the trace of the dial axis on the film is cut by a normal to it from c, then cM' ($=\epsilon\cdot$ sin A; assuming ϵ is represented by $O'c$) is a measure of the dial error and $O'M'$($=\epsilon\cdot$ cos A) of the horizontal arc error.

Figure 40 shows two diffraction patterns of an orthorhombic crystal which permit ready visualization of precession orientation problems. Examination of these yields the following results:[4]

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[1] Fisher, D. J. (1952), *X*-ray precession techniques: *Am. Mineral.*, **37**, 1036–1054. On p. 1050, line 1.; $D=c$ should read $D=c^*$. In line 2 of fn. 7 on p. 1044, V_{vl} should read V_{xl}. On p. 1010 of the preceding paper, line 15 from base, *and* should read *end.*

[2] The numbers here assigned to the figures and tables are in continuation of those used in the paper of footnote no. 1.

[3] Fig. 39 is based on the data of Table 13. These were taken from Table 12 (page 1045) as $\delta-r$ values. It is clear from Fig. 34*b* (page 1046) that $s'c=s'v-cv=\delta-r$; on an instrument for which O' and s' do not coincide, one should use $O'c$ rather than $s'c$.

[4] The H-arc and dial values given are readily obtained using an ordinary slide rule. Set 90° on the sin scale against the ϵ value on the A (or D) scale. Then opposite the value for angle A on the sin scale read the dial correction, and opposite the value for (90°-A) read the H-arc correction.

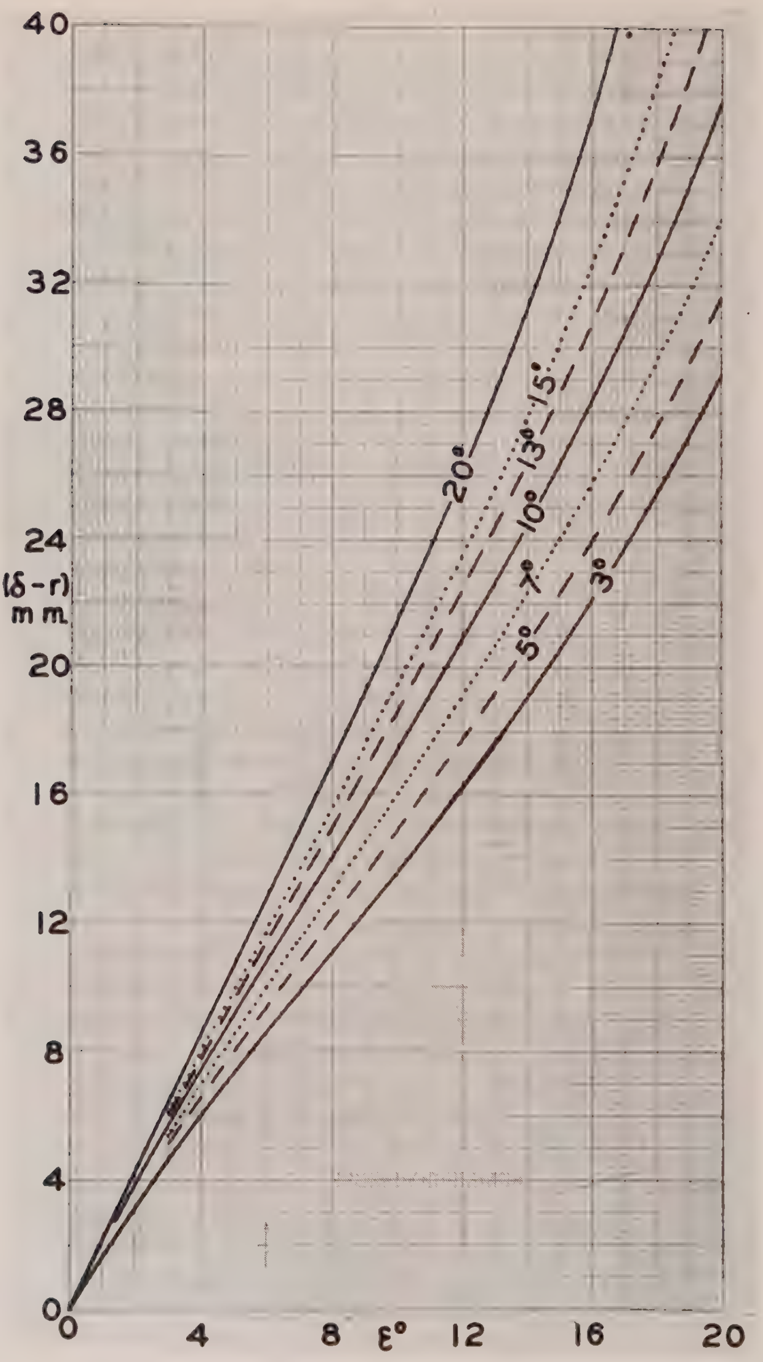

FIG. 39. Graph showing variation in $(\delta - r)$ in mm. with ϵ (in degrees) for various values of $\bar{\mu}$ (the figures on the curves) where F = 60.00 mm.

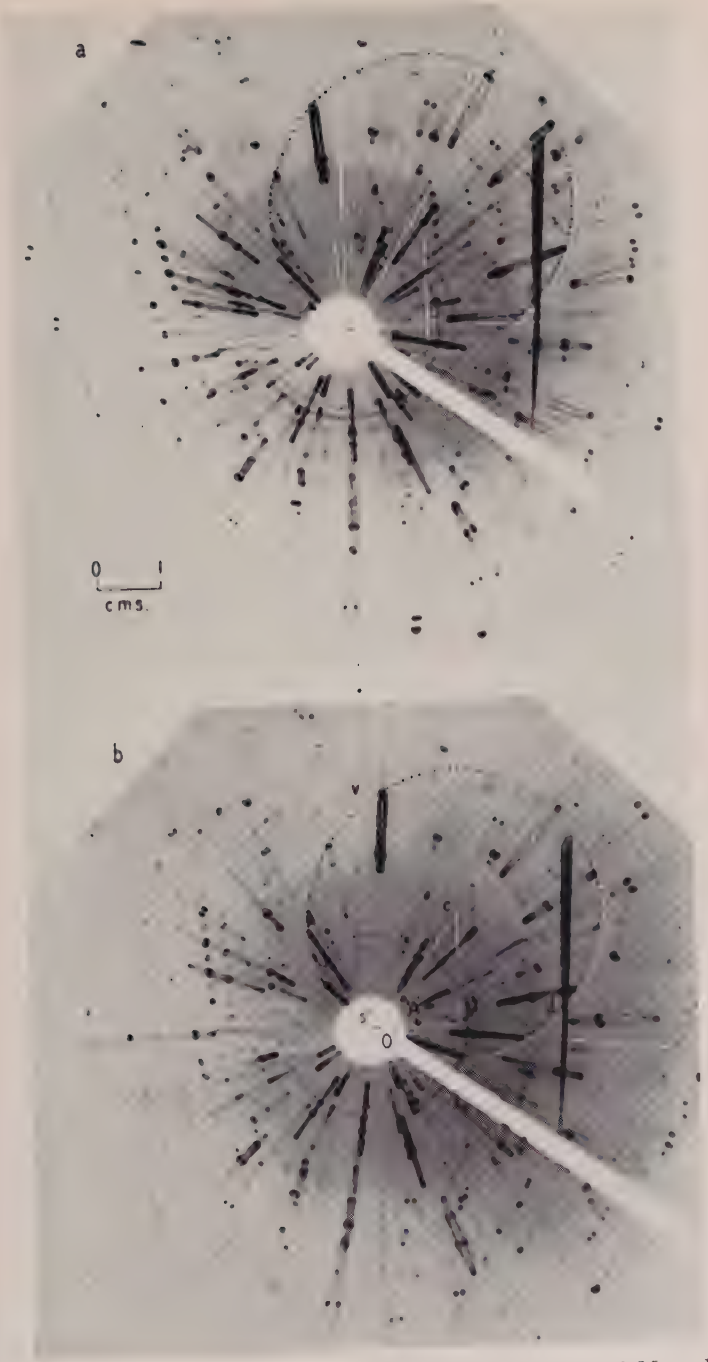

FIG. 40. Precession orientation photos of hopeite with $\bar{\mu}=3°$ and Mo radiation at F = 60.00 mm. If correctly oriented, precession would be on [*a*], with [*c*] parallel to the dial axis. Here the dial reading is off 15° (up), the H-arc off 10° (to the right), and the V-arc off 5°; this last error is up in (*a*) and down in (*b*).

Fig.	Measurements		ϵ	H-arc	Dial
	$O'c$	A	(Fig. 39)	$\epsilon \cdot \cos$ A	$\epsilon \cdot \sin$ A
40*a*	26.1 mm.	58°	18°4	9°8	15°6
40*b*	24.8 mm.	54°	17°6	10°3	14°2

It will be noted that the average values for the dial and H-arc errors are within 0°1 of the true figures. But if one were working from a single picture, the common case, results following this scheme might be as much as 1° off. This is largely because of the error in the orientation of the vertical arc, which prevents this simplified method from being exact. To correct the V-arc setting, one may use the technique given on page 1040. However this is highly empirical, and will not work out as satisfactorily in many cases as one might gather from the example given.

Thus in the orthogonal case of Fig. 40, the proper V-arc setting may be obtained from the bearings of the streak which would be vertical (given correct orientation) and the one that should be horizontal (parallel the dial axis). The bearings of these are as follows:

Fig. 40*a*—N. 6° W. & N. 87° E. (off 6° and 3° resp.)
Fig. 40*b*—N. 4° E. & S. 84° E. (off 4° and 6° resp.)

If one takes the mean errors, these are 4½° and 5° respectively. These mean results are quite close to the true value of 5°. However in the case of an unknown material having a non-orthogonal relationship, this mean error method cannot be applied. It should also be noted that the technique of correcting the V-arc error where c does not coincide with O' (given as s' on page 1040) would be only moderately satisfactory in the case of Fig. 40*a* and would be downright erroneous in the case of Fig. 40*b* (where the so-called correction factor should be negative; the first sentence of the paragraph starting near the top of page 1040 is thus seen to be incorrect).

In conclusion, with a badly misoriented photo[5] such as those in Fig. 40, approximate corrections are computed by the method given above, or by the technique described on page 1039 neglecting the correction factors (of course allowance should be made for the case where an arc is not horizontal). From the next (second) picture, which should be taken with one arc horizontal, it is easy to get the correct setting for this arc from the δ_a-value (measured from O'; not from s'

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[5] It is interesting to note that on such photos the circle of precession becomes slightly elliptical, with the major axis along $O'c$. This contrasts with the case of less severe misorientation, where certain Laue streaks are shortened on the direct beam side of c; see page 1039.

as given on page 1039) and Fig. 31 on page 1041. Then rotate the dial 90° and take a third picture to obtain the correct setting for the other arc. Using these two pictures the desired dial settings can be had from the δ_v-values and Fig. 31.

TABLE 13. LENGTHS OF $O'c$ (in mm.) FOR VARIOUS VALUES OF $\bar{\mu}$ AND ϵ (where F = 60.00 mm.)

$\bar{\mu}$ \ ϵ	1°	2°	3°	5°	7°	10°	13°	15°	20°
3°	1.84	3.37	4.75	7.32	9.81	13.59	17.62	20.55	29.18
5°	1.93	3.63	5.19	8.05	10.79	14.90	19.24	22.38	31.66
7°	1.98	3.78	5.45	8.57	11.54	16.05	20.67	24.05	34.09
10°	2.04	3.94	5.75	9.17	12.45	17.39	22.61	26.38	37.74
13°	2.09	4.07	6.01	9.65	13.23	18.65	24.43	28.65	41.54
15°	2.12	4.16	6.12	9.94	13.71	19.46	25.64	30.17	44.24
20°	2.23	4.35	6.47	10.66	14.89	21.50	28.78	34.25	51.95

Addendum

In this series of papers (*Am. Mineral.*, **37,** 1952, page numbers as indicated) the writer has spoken of "enlarged reciprocal lattice translation values" (p. 100) which are measured on the precession films. It was stated (fn. 8, p. 100) that in non-orthogonal cases these were really enlarged "spacing" values between reciprocal lattice *lines;* but nevertheless the symbol t^* was used to designate these (pp. 101, 104, 109), except on pages 1049–1051, where G, H, J, K, M, & N were employed. Reading Terpstra,[6] it has seemed that the term *interval* is more appropriate to describe the distance between the equal-spaced lines in any plane of the non-orthogonal lattice. The matter is summarized in Table 14

TABLE 14. DESIGNATION OF INTERVALS IN THE ENLARGED TRICLINIC RECIPROCAL LATTICE

Orientation (Table 1, p. 83)	Standard	First Permutation	Second Permutation
Precession axis (pp. 104, 1049–1050)	c	a	b
Intervals	$M^* = i^*_{\perp b^* c}$ $N^* = i^*_{\perp a^* c}$	$G^* = i^*_{\perp ac^*}$ $H^* = i^*_{\perp ab^*}$	$J^* = i^*_{\perp bc^*}$ $K^* = i^*_{\perp a^* b}$

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[6] Terpstra, P. (1952), A Thousand and One Questions on Crystallographic Problems, Groningen. See p. 100.

for the case of the reciprocal lattice; corresponding letters without the asterisk may be used for the direct lattice should occasion arise.

It will be noted that these are much the same symbols as used on pp. 1049–1051, except that the capitals are marked with an asterisk, and the t^* is replaced by i^*. The use of the "meaningless" capitals is much simpler for the printer.

Professor M. J. Buerger informs me[7] that nearly all of his precession orientation work is done using a "layer-line" screen with a *circular* (not annular) opening of 12 mm. diameter (ideal $r=5$ mm. with an extra mm. for tolerance) with $s=28.4$ mm. and $\mu=10°$. If my 7 mm. radius screen (fn. 6*a*, p. 1041) is used with a circular opening (actual diameter 17 mm.) at $\mu=10°$, then it will not cut off the ends of the longer radiating Laue streaks if the orientation error is not greater than 1° with s at 39.7 mm. With a larger error in orientation, the screen should be moved closer to the crystal; thus with the screen at $s=31.7$ mm., the orientation error may be as much as 3° without decapitating the longer Laue streaks. This method of taking orientation pictures is very rapid and yields "clean" negatives that lack the sometimes-confusing n-level Laue streaks.

Erratum

Assuming the negative ends of the a and b axes are where indicated in Fig. 36 (p. 1049), then $+c^*$ must extend *up* normal to the paper (fn. 8 on p. 1048 should be corrected). This means that $+ c^*$ must extend to the *left* in Fig. 37*b*, hence the signs for all the digits of the node indices in this figure must be reversed. Also $+c^*$ should be changed to read $-c^*$ on p. 1051 (line 8 from base) and p. 1054 (line. 2).

[7] Personal communication, Dec. 24, 1952 and Jan. 13, 1953.

TWO DEFINITIONS OF POSITIVE AND NEGATIVE EXTINCTION ANGLES IN THE PLAGIOCLASE FELDSPARS: ONE LEADING TO CONSISTENCY AND CLARITY, THE OTHER TO INCONSISTENCY AND CONFUSION

George Tunell, *University of California at Los Angeles, Los Angeles, California.*

An extinction angle has been defined as positive by Duparc and Reinhard,[1] by Rosenbusch and Mügge,[2] and by Chudoba,[3] if the rotation of the line generating the angle proceeds clockwise from the reference direction (cleavage trace, trace of composition plane of twins, trace of crystal face) to the X'-vibration direction. (Under the microscope one actually turns the stage and crystal with respect to the line, the North-South crosshair, which remains stationary. Thus, if one first sets the trace of the 010-cleavage to coincide with the North-South crosshair and then rotates the stage counterclockwise through an angle less than 90° to a position of extinction, and if the vibration direction which then coincides with the North-South crosshair is the X'-vibration direction, the sign of the extinction angle is positive.) Stereographic projections of the extinction angles of the plagioclase feldspars with the trace of the 010–cleavage (or composition plane) as the reference direction have been published in several books.[4] Each of these projections gives the extinction angles for the upper half of the crystal. Of course, the numerical value of the extinction angle of any section of a crystal will be the same if the section is turned over. Hence from the numerical values of the extinction angles for the end of the crystal represented in the projection, one can obtain the numerical values of the extinction angles for the other end. The convention that an extinction angle is considered positive if the line generating the angle is turned clockwise from the reference direction to the X'-vibration direction necessitates that the sign of the extinction angle is reversed when the section is turned over. In conformity with this requirement, in the published stereographic projections the extinction angles at points on the circumference of the projection at opposite ends of a diameter have opposite signs (cf. Fig. 1); these two extinction angles, numerically equal, but opposite in sign, of course apply to the same section; the first extinction angle applies with one side of the section up, the second, numerically equal, but opposite in sign, applies when the section is turned over.

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[1] Duparc, L., and Reinhard, M., La détermination des plagioclases dans les coupes minces: *Mémoires de la Société de Physique et d'Histoire Naturelle de Genève*, **40,** 14, 22 (1924). (It is to be noted that the statement of the definition of a positive extinction angle by Duparc and Reinhard was not accompanied by any qualification limiting the application of the definition to the upper half of the crystal.)

The statement was made by Duparc and Reinhard and by Rosenbusch and Mügge in the publications cited that in sections perpendicular to the *a*-axis the sign of the extinction angle is positive if the X'-vibration direction lies in the acute angle between the traces of the two perfect cleavages; to this these authors added the statement that in the case of sections perpendicular to the *a*-axis the sign of the extinction angle

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[2] Rosenbusch, H.-Mügge, O., Mikroskopische Physiographie der petrographisch wichtigen Mineralien, Bd. I, 2$^{\text{te}}$ Hälfte, 5$^{\text{te}}$ Aufl., S. 745, Schweizerbart'sche Verlagsbuchhandlung, Stuttgart (1927).

[3] Chudoba, K., Die Feldspäte und ihre praktische Bestimmung, S. 13, Schweizerbart'sche Verlagsbuchhandlung, Stuttgart (1932).

[4] Cf. Duparc and Reinhard, *op. cit.*, Plates 1–8, and Rosenbusch-Mügge, *op. cit.*, Plates 30–35.

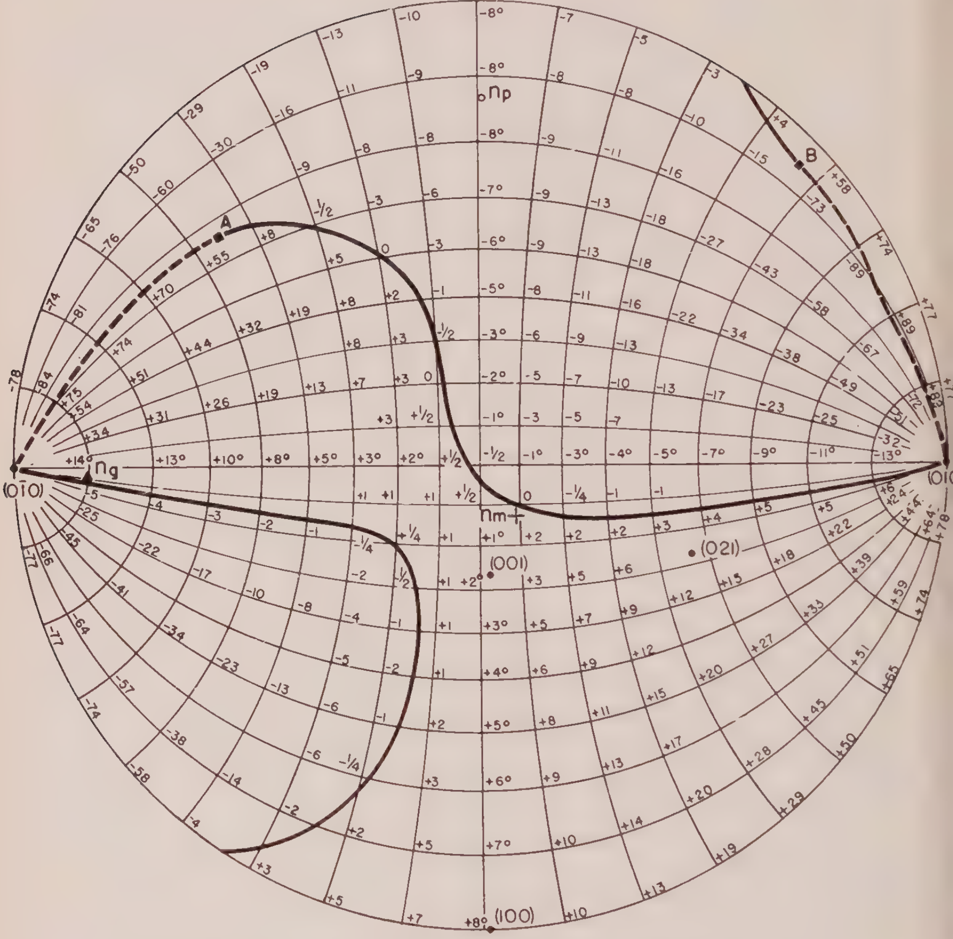

FIG. 1. Stereographic projection of the extinction angles of oligoclase from Duparc and Reinhard's "La Détermination des Plagioklases dans les Coupes Minces."

can practically always be determined, since the two perfect cleavages are practically always sufficiently developed so that their traces can be observed, but that in the cases of sections parallel to 001 or perpendicular to a bisectrix, the sign of the extinction angle usually cannot be determined.[5] Now the statement that the sign of the extinction angle of a section perpendicular to the *a*-axis is positive when the X'-vibration direction lies in the acute angle between the traces of the two perfect cleavages conflicts with the original convention that an extinction angle is positive if the line generating the angle is turned clockwise from the reference direction to the X'-vibration direction. The acute angle forms the upper left corner of the section perpendicular to the a-axis when it is viewed from the positive end of the a-axis, and the upper right corner when it is viewed from the negative end of the a-axis. Thus with a section perpendicular to the a-axis, angles are turned counter-clockwise from the trace of the 010-cleavage (or composition plane) to a line lying in the acute angle between the traces of the perfect cleavages if one is looking at the section from the positive end of the a-axis, and clockwise if one is looking at the section from the negative end of the a-axis. The convention that with a particular section of a crystal the sign of the extinction angle is the same no matter which side of the section is toward the observer, requires that at a point on the projection sphere near some great circle, the sign of the extinction angle be the opposite of the sign of the extinction angle at a second point on the other side of and close to the same great circle, although the extinction angle does not pass through zero between these two points, a very unsatisfactory outcome. A convention that the sign of the extinction angle is the same no matter which side of the section is toward the observer requires that positive extinction angles be turned clockwise from the reference direction to the X'-vibration direction on one half of the crystal and counter-clockwise on the other half, a very unnatural and arbitrary rule. Moreover, with this convention along one half of the great circle arbitrarily chosen to divide the sphere of projection into two halves, positive extinction angles must be turned clockwise from the reference direction to the X'-vibration direction, and along the other half of this great circle, positive extinction angles must be turned counter-clockwise from the reference direction to the X'-vibration direction. In the stereographic projections of Duparc and Reinhard, and those of Rosenbusch and Mügge, however, the signs of the extinction angles at the opposite ends of any diameter of the great circle forming the circumference of the projection are not the same. Thus, Duparc and Reinhard, and Rosenbusch and Mügge did not adhere consistently to a convention that the sign of the extinction angle is the same no matter which side of the section is toward the observer. On the other hand, neither did they adhere consistently to the alternative convention that the sign of the extinction angle is positive if the line generating the angle is turned clockwise from the reference direction to the X'-vibration direction no matter which side of the section is toward the observer.

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[5] Chudoba made equivalent statements in the cases of sections perpendicular to 010 and sections perpendicular to the *a*-axis. *Op. cit.*, pp. 13, 23, 24.

If the original convention that an extinction angle is positive when the line generating the angle is turned clockwise from the reference direction to the X'-vibration direction be applied to all sections of the crystal, then along any circle on the projection sphere the sign of the extinction angle will only change from positive to negative as the value of the extinction angle passes through zero. This convention is therefore a satisfactory one and does not lead to any inconsistency. It should be noted also that application of the original convention to both ends of a crystal enables one to determine the sign of every observed extinction angle even in those cases in which one cannot determine from which side he is viewing the section. With this definition of the sign of the extinction angle the published curves representing the extinction angles of the plagioclase feldspars for certain sections, or the maximum extinction angle in certain zones, presuppose that a particular side of each section is toward the observer. The curve for the other side is obtained by changing the sign of each value, or, in other words, by reflecting the original curve across the axis of abscissae. Thus, for example, the curve of the extinction angles in sections perpendicular to the *a*-axis viewed from the negative end of the *a*-axis is represented in Fig. *2a*. The curve of the extinction angles in sections perpendicular to the *a*-axis viewed from the positive end of the *a*-axis is represented in Fig. *2b*. With plagioclase feldspars there is no uncertainty as to which end of the *a*-axis this section is viewed from, because the angle between the traces of the two perfect cleavages in the plane of the section is perceptibly different from 90°. Therefore, one knows whether the value of the extinction angle that he measures is to be compared with the curve of Fig. *2a* or with that of Fig. *2b*. In the case of the maximum extinction angle in the zone perpendicular to 010 the curve given in numerous books is for the top half of the crystal; the curve for the bottom half is again obtained by changing the sign of each value, or, in other words, by reflecting the original curve across the axis of abscissae. In Fig. 3 the solid curve represents the maximal extinction angles for the zone perpendicular to 010 for the top half of the crystal and the dotted curve represents the maximal extinction angles for the same zone for the bottom half of the crystal. When an albite twin is found showing the maximum extinction angle of all the sections in the zone perpendicular to 010, the determination of the composition of the plagioclase by the use of these curves is unique, except in cases in which the maximum extinction angle is less than 17°. If the maximum extinction angle is less than 17°, and if the sign of the extinction angle is positive, the intersections of the horizontal line representing the observed positive maximum extinction angle with the two curves for the two ends of the crystal give two possible compositions of the plagioclase. Since it is not possible to prove from which side the section was viewed in this case, it is not possible to tell which of the

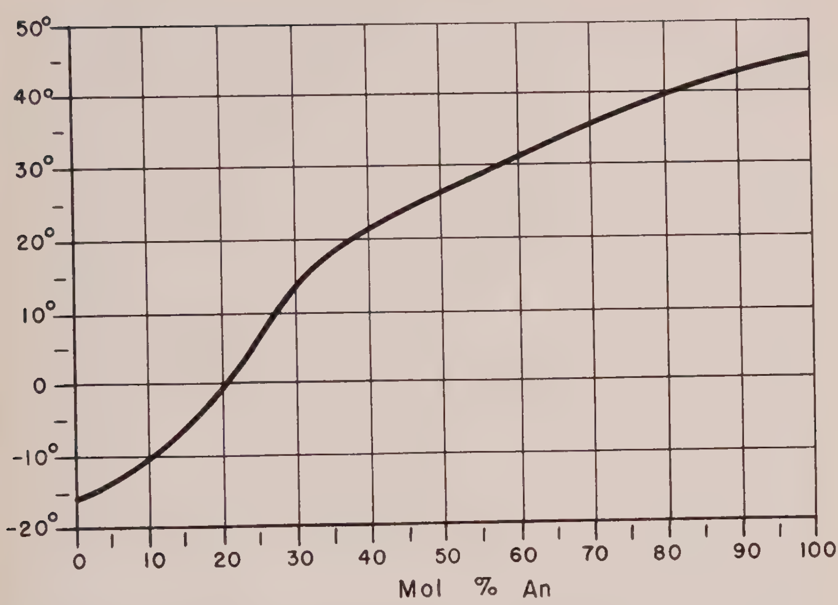

FIG. 2*a*. Extinction angles of the plagioclases in sections perpendicular to the *a*-axis viewed from the negative end of the *a*-axis.

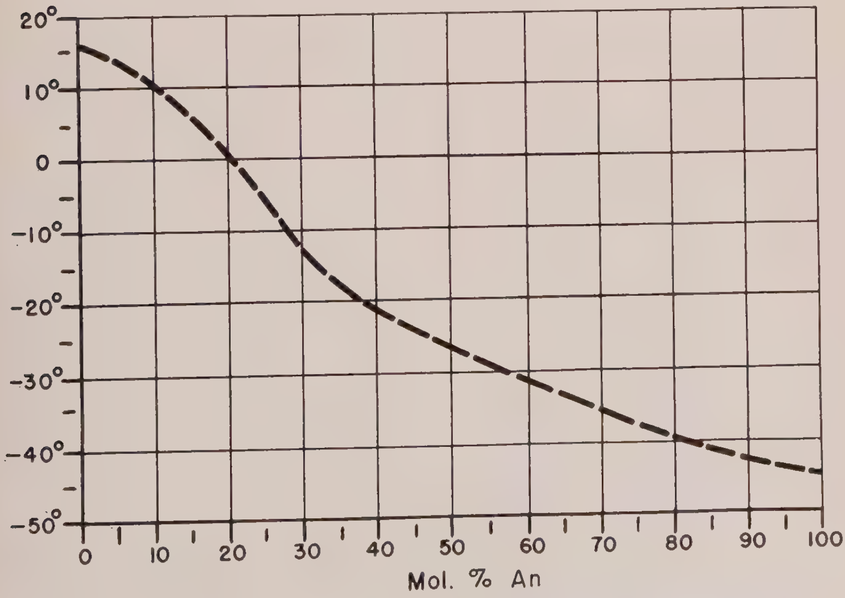

FIG. 2*b*. Extinction angles of the plagioclases in sections perpendicular to the *a*-axis viewed from the positive end of the *a*-axis.

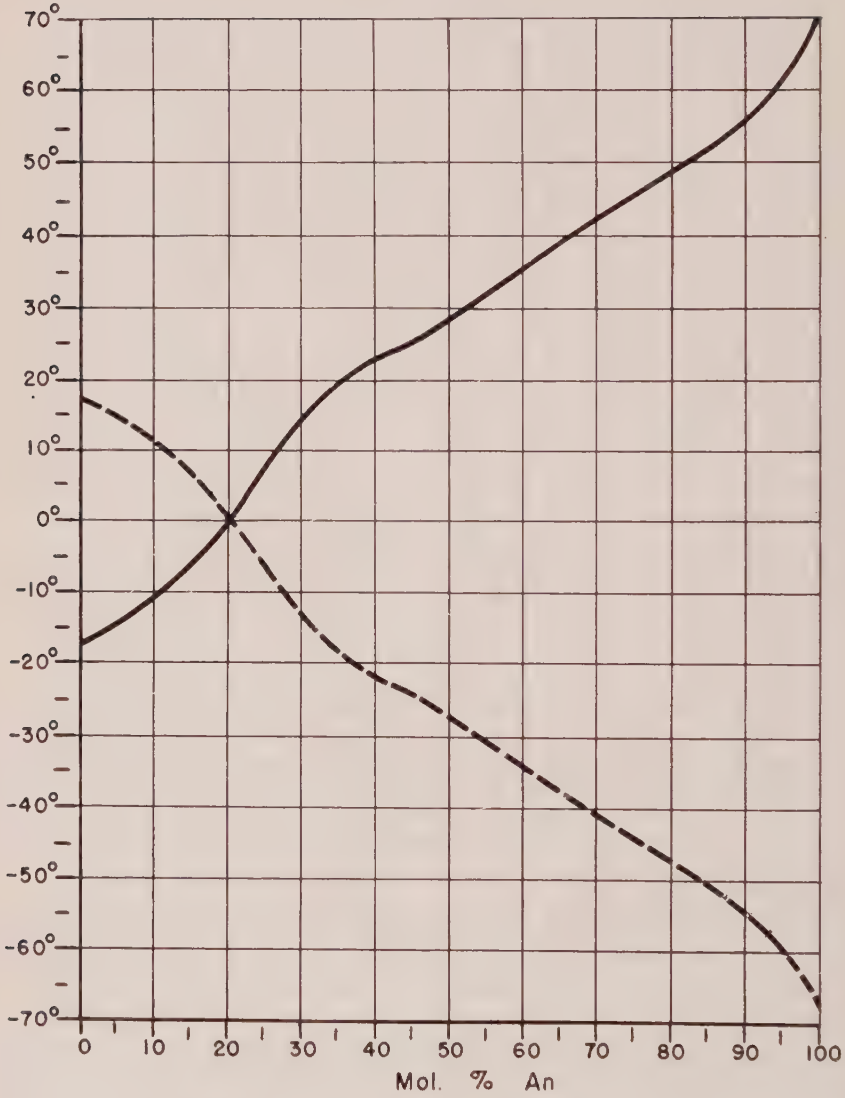

FIG. 3. Maximal extinction angles of the plagioclases in the zone perpendicular to the 010-pinacoid.

two alternative compositions is the correct one by the use of this method alone. If the sign of the maximum extinction angle were negative, the numerical value being the same as in the preceding case, the intersections of the horizontal line representing the observed negative maximum extinction angle with the two curves for the two ends of the crystal would again give two possible compositions, and these two possible compositions would be the same ones that were found in the preceding case.

CONCLUSION

The confusion caused by the simultaneous use of two conflicting conventions for the sign of the extinction angle can and should be resolved by adherence to the one convention that the sign of the extinction angle is positive if the line generating the angle is turned clockwise from the reference direction to the X'-vibration direction no matter which end of the crystal is being viewed. (It must always be remembered, of course, that under the microscope one actually turns the stage and crystal with respect to the line, the North-South crosshair, which remains stationary. Thus if one first sets the trace of the 010–cleavage to coincide with the North-South crosshair and then rotates the stage counter-clockwise through an angle less than 90° to a position of extinction, and if the vibration direction which then coincides with the North-South crosshair is the X'-vibration direction, the sign of the extinction angle is positive.) This convention leads to simple, straightforward technique and to consistent statements of all the optical relations.

STUDIES OF BORATE MINERALS: 1—*X*-RAY CRYSTALLOGRAPHY OF COLEMANITE[1]

C. L. CHRIST, *U. S. Geological Survey, Washington, D. C.*

As part of a general *x*-ray study of borate minerals begun in the U. S. Geological Survey laboratories, the crystal structure of the most common calcium borate, colemanite, is being investigated. This investigation has begun with a redetermination of the space group and of precise lattice constants for colemanite. Powder patterns using Cu and Cr radiation have been measured and indexed.

The crystals of colemanite used came from Meyerhofferite Tunnel in Twenty Mule Team Canyon, Death Valley, Inyo County, Calif.; they were furnished by Waldemar T. Schaller, U. S. Geological Survey. The crystals used are small, colorless, transparent, prismatic elongated along [001], with large (110).

Weissenberg patterns, using Cu radiation, were prepared for the zero and upper levels around [001] and [010]. The systematic extinctions observed, namely that reflections $(0k0)$ with k even only, and $(h0l)$

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[1] Publication authorized by the Director, U. S. Geological Survey.

with h even only are present, lead unequivocally to the space group $P2_1/a$.

For the purpose of obtaining precise lattice constants, Weissenberg patterns of small crystals (about 0.2×0.2×0.4 mm.) around [001] and [010] (Cu radiation) with superimposed powder patterns of silver were prepared. The powder pattern of silver (of precisely known lattice constant) on each colemanite film that was measured enabled a precise camera constant to be assigned to that film. The corrected d-spacings thus obtained from the films were then plotted as a function of $\sin^2 \theta$ and the extrapolated values determined in the usual fashion (Bradley and Jay, 1932). A full discussion of the use of the conventional Weissenberg camera in measuring lattice constants of single crystals precisely, with colemanite as an example, will be discussed elsewhere (Christ, in preparation).

The single crystal data for colemanite are collected in Table 1.

TABLE 1. SINGLE CRYSTAL DATA: COLEMANITE—$2CaO \cdot 3B_2O_3 \cdot 5H_2O$

Space Group—$P2_1/a(C^5_{2h})$

$a=8.743 \pm 0.004$ Å	Z=2
$b=11.264 \pm 0.002$ Å	density (calc.)=2.419
$c=6.102 \pm 0.003$ Å	density (meas.)=2.42
$\beta=110°7' \pm 5'$	V=564.2 Å^3
$a:b:c=0.7762:1:0.5418$	

(Values based on Cu radiation: $\lambda K\alpha_2=1.54434$, $\lambda K\alpha_1=1.54050$, $\lambda K_\beta=1.39217$ Å; Ag: a=4.0861 Å)

Two previous x-ray studies have been made on colemanite by Dér (1941) and Nikolsky (1940). Their results are compared with those obtained in the present study in Table 2.

TABLE 2. COMPARISON OF X-RAY RESULTS

	a	b	c	β	Space Group
Present work	8.743	11.264	6.102	110°7′	$P2_1/a$
Dér (1941)	8.63	11.14	6.11	110°9′*	$P2_1/a$
Nikolsky (1940)	8.74	11.31	6.07	110°9′*	$P2/m$

The lattice constants are given in Å units. The values originally given by Dér and Nikolsky are assumed to be in kX units and are recalculated to Å units above.

* Assumed by Dér and Nikolsky from the morphological measurements.

It may be remarked that the Weissenberg patterns prepared in the present study clearly show the extinctions leading to $P2_1/a$, in agree-

TABLE 3. X-RAY POWDER DATA: COLEMANITE—$2CaO \cdot 3B_2O_3 \cdot 5H_2O$

Monoclinic $P2_{1/a}$; $a=8.743$, $b=11.264$, $c=6.102$ Å
$\beta=110;7'$, $Z=2$

Measured*		Measured*		Calculated	
Cu/Ni	$\lambda=1.5418$ Å	Cr/V	$\lambda=2.2909$ Å		
I	d_{hkl}	I	d_{hkl}	d_{hkl}	hkl
4	6.66		6.69	6.634	110
		wk.	5.74	5.730	001
50	5.64		5.64	5.632	020
18	5.10		5.09	5.107	011
				5.093	$11\bar{1}$
4	4.65		4.65	4.644	120
36	4.00		4.00	4.017	021
				4.010	$12\bar{1}$
50	3.85		3.85	3.857	210
				3.840	111
36	3.29		3.29	3.295	$22\bar{1}$
100	3.13		3.13	3.137	$13\bar{1}$
				3.141	031
36	2.894		2.895	2.898	201
25	2.805		2.806†	2.807	211
				2.795	$31\bar{1}$
25	2.772		2.761	2.770	230
				2.768	$21\bar{2}$
				2.764	131
				2.758	$23\bar{1}$
9	2.660		2.659	2.663	140
				2.659	310
50	2.550		2.546	2.554	022
				2.547	$22\bar{2}$
18	2.392		2.390	2.399	112
				2.387	$31\bar{2}$
18	2.315		2.314	2.318	141
				2.315	$24\bar{1}$
9	2.196		2.207	2.211	330
2	2.176		2.168	2.172	150
36	2.141		2.142	2.145	$41\bar{1}$
18	2.094		2.090	2.097	051
				2.096	$15\bar{1}$
18	2.067		2.070	2.067	$14\bar{2}$
18	2.045		2.043	2.047	$33\bar{2}$
				2.044	321
				2.043	202
50	2.010		2.011	2.015	$34\bar{1}$
				2.010	212
				2.008	042

TABLE 3—(*Continued*)

Measured*		Measured*		Calculated	
Cu/Ni	λ=1.5418 Å	Cr/V	λ=2.2909 Å		
I	d_{hkl}	*I*	d_{hkl}	d_{hkl}	*hkl*
25	1.969		1.969	1.972 1.970	151 $25\bar{1}$
18	1.892		1.891	1.894	331
		v wk.	1.887	1.889	$43\bar{1}$
3	1.848		1.848	1.851 1.845	142 $34\bar{2}$
3	1.814		1.809	1.812 1.809	$15\bar{2}$ 023
9	1.778		1.781	1.779 1.779 1.777	$13\bar{3}$ 251 $23\bar{3}$
9	1.728	Plus additional weak lines			
9	1.701				
3	1.656				
3	1.627				
9	1.569				
2	1.545				
4	1.481				
4	1.453				
4	1.383				
4	1.344				

Plus additional weak lines.

* Patterns corrected for shrinkage.

† Doublet.

ment with Dér, and in disagreement with Nikolsky. The space group listed in Dana (1951, p. 349) is incorrect.

The morphology of colemanite has been studied by several investigators; the results of these investigations are summarized by Hutchinson (1912). It suffices here to compare the axial ratio and value of the monoclinic angle obtained in the present study with those obtained by Eakle (1902) and by Jackson (1885) from morphological measurements. The values given by other investigators all lie between those given by Eakle and Jackson.

Present study: $a:b:c=0.7762:1:0.5418$, $\beta=110°7'$
Eakle: $a:b:c=0.7768:1:0.5430$, $\beta=110°7'$
Jackson: $a:b:c=0.7748:1:0.5410$, $\beta=110°9'$

The x-ray powder data for colemanite are collected in Table 3.

In regard to the structure of colemanite it is interesting to note that the calculated volume per oxygen atom is 17.6 Å^3. This is about the value to be expected if the structure is essentially determined by the nearly close-packing of oxygen ions with the boron and calcium ions occupying interstices in the oxygen framework. The perfect cleavage of colemanite parallel to (010) indicates that the oxygen ions are linked by boron ions in sheets parallel to (010). Work on the structure is proceeding.

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PEACOCK MEMORIAL PRIZE,* 1953

offered by the

WALKER MINERALOGICAL CLUB

Royal Ontario Museum of Geology and Mineralogy
100 Queen's Park, Toronto 5

THIS PRIZE OF ONE HUNDRED DOLLARS cash is again offered for the best scientific paper on pure or applied mineralogy (including crystallography and petrology) submitted to the Club by a graduate student at any university or similar institution.

Conditions:

1. The paper submitted may offer new or refined observations, or a significant synthesis and interpretation of existing data, or some new or improved application of mineralogy to useful ends, or the results of other work of sufficient interest and value.
2. The papers should be typed double-spaced on letter-size paper and illustrations should be carefully prepared and suitable for reproduction. References to the literature should follow the style in *Contributions to Canadian Mineralogy* as printed in *The American Mineralogist* (May–June issues, 1949–50–51–52).
3. Each paper must be accompanied by a letter from the student's supervisor stating the nature and extent of the assistance he may have given to the work submitted.
4. The papers are to be sent to the Secretary, Walker Mineralogical Club, 100 Queen's Park, Toronto, *not later than June 30, 1953*. They will be judged by a committee appointed by the Council of the Club who will consider both the value of the papers and their suitability for publication without substantial changes. If no paper of sufficient merit is received, the prize will not be given.
5. The Committee will select one or more of the papers for presentation by the authors or other designated persons at a regular meeting of the Club, and the result of the contest will be announced at that time.
6. The prize winning paper, and perhaps others of sufficient merit, will be published in *Contributions to Canadian Mineralogy.*

* Instituted in 1947 as the Walker Mineralogical Club prize, was renamed the Peacock Memorial Prize in 1950 in honour of the late Professor Martin A. Peacock, President of the Walker Mineralogical Club 1942–43, and Editor of Contributions to Canadian Mineralogy, 1944–50.

Prizes have been awarded to:

- A. R. GRAHAM, Queen's University (1947), "*X*-ray study of chalcosiderite and turquoise": *Univ. Toronto Studies, Geol. Series*, No. **52,** 1948 (*Contributions to Canadian Mineralogy*, Vol. **4**, No. 4).
- I. H. MILNE, Univ. of Toronto (1948), "Chloritoid from Megantic County, Quebec": *Am. Mineral.*, **34,** 1949 (*Contributions to Canadian Mineralogy*, Vol. **5,** No. 1).
- R. J. ARNOTT, University of Manitoba (1950), "*X*-ray diffraction data on some radioactive oxide minerals": *Am. Mineral.*, **35,** 1950 (*Contributions to Canadian Mineralogy*, Vol. **5,** No. 2).
- R. J. TRAILL, Queen's Univ. (1952), "Synthesis and *x*-ray study of uranium sulphate minerals": *Am. Mineral.*, **37,** 1952. (*Contributions to Canadian Mineralogy*, Vol. **5,** No. 4).

ANNUAL MEETING

The thirty-fourth annual meeting of the Mineralogical Society of America will be held in Toronto, Canada, Monday through Wednesday, November 9–11, 1953. A series of field trips is being planned for Thursday, Friday and Saturday following the meeting.

Abstracts of papers to be presented at the annual meeting must be received by the Secretary on or before *July 15, 1953*. Abstracts blanks may be obtained from the Secretary.

C. S. HURLBUT, JR., *Secretary*

INTERNATIONAL UNION OF CRYSTALLOGRAPHY

The Executive Committee has accepted a kind invitation from the French Government to hold the Third General Assembly and International Congress in Paris from 21 to 28 July 1954. These dates have been chosen in consultation with the French National Committee for Crystallography and with the National Committees of all the Adhering Bodies. It is hoped that this early notice will make it possible for crystallographers throughout the world to arrange to attend.

Further particulars of the meeting will be given from time to time in *Acta Crystallographica.*

The Geology Department of the School of Mines and Metallurgy of the University of Missouri, Rolla, Missouri, wishes to announce the availability of several graduate assistantships and research fellowships for the academic year 1953–54. Application blanks and information concerning these may be obtained from Dr. O. R. Grawe, Chairman of the Geology Department.

It is announced that Dr. Robert W. Webb has accepted the joint position of Executive Director of the American Geological Institute and Executive Secretary of the Division of Geology and Geography of the National Research Council, effective January 1, 1953.

Dr. Webb will be on leave of absence from his position as Professor of Geology, Department of Physical Science, Santa Barbara College, University of California.

A.G.I. Report No. 8, "*Departments of Geological Science in Educational Institutions of the U. S. and Canada,*" *December 1952* (revision of Report #6, March 1952), has been sent to the printer and will be available for distribution in either January or February, 1953, depending upon the number and nature of crises that develop in its reproduction by a direct offset process. Price, $1.00. Orders now being held will be filled at the first possible moment. Address: American Geological Institute, 2101 Constitution Ave., Washington, D. C.

Copies of Resource Appraisals Map MR-1, "Geologic Environment of Alumina Resources of the Columbia River Basin" can be obtained at 80 cents each from the Chief of Distribution, Geological Survey, Department of the Interior, Washington 25, D. C., for requests originating east of the Mississippi River, or from the Geological Survey, Rocky Mt. Service and Supply, Denver Federal Center, Bldg. 25, Colo., for requests originating west of the river.

"Table 1 gives the chemical analyses of potential alumina ores in the Basin and vicinity. Table 2 shows the reserves of potential alumina ore in millions of short tons (dry). . . . Other potential recoverables from the Columbia Basin aluminous ores include titanium, iron, and gallium. . . . An extensive list and bibliography provide details on the known deposits by subject, county, and author."

The American Association of Spectrographers is planning a symposium on "Emission Spectroscopic Determination of Metals in Non-Metallic Samples." The symposium will be held in Chicago on May 1, 1953. Contributed papers in the fields of petroleum, geology, agriculture, pharmacy, biology, ceramics, etc., are welcome. Please address all inquiries to:

J. P. Pagliassotti, *Chairman;* Symposium Committee,
c/o Standard Oil Company (Indiana),
Box 431, Whiting, Indiana

The Society for Experimental Stress Analysis will hold its Spring Meeting on May 20, 21, and 22, 1953 at Hotel Schroeder, Milwaukee, Wisconsin. Inquiries may be directed to Mr. W. F. Hofmeister, Metallurgical Laboratory, Chain Belt Company, 4501 West Greenfield Ave., Milwaukee, Wisconsin.

REPORT OF THE COMMITTEE ON THE MEASUREMENT OF GEOLOGIC TIME, 1951–52

The 1951–52 report of the Committee on the Measurement of Geologic Time, of the Division of Geology and Geography of the National Research Council of the National Academy of Sciences, published in January 1953, is available on order to the Division, 2101 Constitution Avenue, N.W., Washington 25, D.C., price: $1.50. This is the latest in a series of reports that have been in great demand since the Committee was formed in 1924.

OFFICERS: AMERICAN GEOLOGICAL INSTITUTE

President: Dr. Carey Croneis, President, Beloit College, Beloit, Wisconsin.
Vice Pres.: Dr. W. C. Krumbein, Professor of Geology, Northwestern University, Evanston, Illinois.
Sec'y-Treas.: Dr. Harry S. Ladd, U. S. Geological Survey, Washington 25, D. C.
Executive Director: Dr. Robert W. Webb, 2101 Constitution Avenue, N.W., Washington 25, D. C.

All communications regarding the *News Letter* and the Institute may be addressed to the Executive Director. Suggestions, contributions, and criticisms are welcome.

BOOK REVIEWS

THE SYSTEM OF MINERALOGY OF JAMES DWIGHT DANA AND EDWARD SALISBURY DANA, Yale University, 1837–1892; seventh edition, entirely rewritten and greatly enlarged, by CHARLES PALACHE, HARRY BERMAN, AND CLIFFORD FRONDEL, Harvard University; Volume II, halides, carbonates, nitrates, iodates, borates, sulfates, selenates, tellurates, chromates, phosphates, arsenates, vanadates, antimonates, vanadium oxysalts, molybdates, tungstates and organic compounds. New York (John Wiley and Sons), London (Chapman and Hall) 1951, xi+1124 pages, many figures $6\frac{1}{4}\times 9\frac{1}{4}$ inches. Price, $15.00.

It is a pleasure to note the appearance of the second volume of the work that has dominated American mineralogy for well over one hundred years and has been very prominent among mineralogical works the world over during much of this time.

In his excellent review of Volume I, Peacock (*Am. Mineral.*, **29**, 453–454, 1944) discusses most of the ways in which the Seventh Edition differs from and extends the scope of the Sixth. There is no point to repeating this discussion, and not a great deal could be added to it. However, Peacock did not mention that the new edition includes optical properties of opaque minerals. This is a big step forward; it will greatly increase the usefulness of the work and will be welcomed by all mineralogists.

The new edition will no doubt maintain the position of its predecessor. The amount of compilation, original work, and editorial effort that has gone into this revision is almost beyond comprehension, especially when it is realized that the brunt of the preparation of Volume II has been borne by a single individual.

The reviewer regards this tremendous undertaking with a certain amount of awe, but he cannot agree with Spencer (*Min. Abs.*, **11**, 463, 1952) that "Criticism . . . would be . . . sacrilegious." Any work of this magnitude must treat of subjects on which there are legitimate differences of opinion and must be subject to improvement in the manner of presentation of some of the material. Attention is respectfully called, therefore, to several topics and usages that might be handled somewhat differently to good advantage.

For example, substitution of nO, nE, nX, nY, and nZ for ω, ϵ, α, β, and γ to indicate indices of refraction goes from a simple to a much more complex and potentially ambiguous system. The Greek letters have long been used in this connection and their meaning is clear and unmistakable. They are in widespread use in the analytical expressions of the theory of crystal optics. In many of these equations the letters X, Y, and Z are used as space co-ordinates. A glance at such equations in Johannsen's Manual of Petrographic Methods, Wahlstrom's Optical Crystallography, the works of F. E. Wright, or similar treatments will show that the use of α, β, and γ for indices of refraction is much more pleasing and less confusing than substitution of nX, nY, and nZ would be. For example, the well-known equation

$$\frac{X^2}{\alpha^2}+\frac{Y^2}{\beta^2}+\frac{Z^2}{\gamma^2}=1$$

becomes

$$\frac{X^2}{nX^2}+\frac{Y^2}{nY^2}+\frac{Z^2}{nZ^2}=1$$

and derivations from this and related equations are even more complicated and confusing. It would be much simpler and more logical to continue to use ω, ϵ, α, β, γ and X, Y, Z with their classical meanings of indices of refraction and space co-ordinates, respectively, than to attempt to change either or both usages.

It is unfortunate that changes were made in the letters used to designate the various forms of calcite. Admittedly, the system proposed by Goldschmidt in 1886 has priority over the usage of Dana. However, the letters used in the Sixth Edition have been so widely employed during the past thirty years in studies of calcite-bearing tectonites and of artificial deformation of marbles that great confusion would result now by a change, for example, to δ instead of e to represent $\{01\bar{1}2\}$ or to p instead of r to represent $\{10\bar{1}1\}$. Moreover, the new edition does not follow the older scheme exclusively; it retains $c\{0001\}$ and $m\{1010\}$, whereas Goldschmidt used o and b for these forms. Also, the new edition employs $e\{0112\}$, $r\{1010\}$, $v\{21\bar{3}1\}$, etc., for the other rhombohedral carbonates, so the usage is not consistent in any respect.

The handling of literature references could have been improved in several ways. In the Sixth Edition the references were given together with the information taken therefrom; in the Seventh Edition the references are given in lists at the end of the description of each species, and it is commonly necessary to turn several pages to find a reference. There is no easy way to find out whether a given citation (except those later than 1915) or an analysis is repeated from the earlier edition; some analyses are repeated in the same form; some are reported in different components so that they must be recalculated to be sure whether they are the same analyses or different ones taken from the same reference. A simple indication such as an asterisk for each analysis new to this edition would greatly simplify reference for the reader.

In Volume I, a complete bibliography was given, including periodicals with their respective abbreviations and other pertinent data, and a list of books consulted. The more than seven years that intervened before Volume II appeared were marked by many changes in mineralogical literature; new books have appeared; new journals, such as the *Heidelberger Beiträge*, *Geochimica et Cosmochimica Acta*, and *Acta Crystallographica* have been established; major changes have taken place in some publications, e.g., in the *Neues Jahrbuch;* and other publications have been discontinued. No mention is made of these changes in Volume II; perhaps the bibliography can be brought up to date when Volume III is published.

In the preface to Volume I, emphasis is placed on "a new method of treating minerals that form a so-called series." This new method is also followed in Volume II but leaves a great deal to be desired. There is not a single diagram representing the properties of a mineral series in either volume; for the most part, properties are not correlated with the analyses given. Properties (indices of refraction) are commonly given for only one intermediate composition, so that one must interpolate for other compositions, and the interpolation is not accurate if the property is not a straight-line function.

All the crystal drawings have been redrawn and are, for the most part, larger than the figures in the older edition. Correlation with the discussion in the text, however, is much more difficult than it was in the previous edition. This is partly because the figures are not numbered and partly because references in the text are to form symbols, whereas the faces on the drawings of crystals are designated by letters.

Several examples of nomenclature might be questioned; one glaring case to which attention should be called is the relegation of the well-established name *martinite* to the synonymy and the substitution therefor of *whitlockite*.

Typographical errors are more numerous than in the Sixth Edition but are neither serious enough to call for a listing nor abundant enough to be noticeable in ordinary reading or reference. With a work of this magnitude that has been so completely rewritten and has had so much and so many kinds of new material added, the surprising thing is that there are not more errors, inconsistencies, and oversights.

Professor Frondel, who is very largely responsible for the preparation of Volume II, is to be heartily congratulated on its appearance and on the fact that it maintains the high standard set by Volume I and by the magnificent Sixth Edition. Volume III, the Silicates, which will complete the monumental Seventh Edition, is eagerly awaited.

Earl Ingerson,
U. S. Geological Survey, Washington 25, D. C.

A THOUSAND AND ONE QUESTIONS ON CRYSTALLOGRAPHIC PROBLEMS,

by P. Terpstra. Publication from the Crystallographic Institute of the University of Groningen. J. B. Wolters, Groningen, Djakarta. 195 pp., 183 figs., 1952. Paper bound.

The author has prepared a collection of problems covering an impressive range of crystallographic subjects. He states that the book is intended for use by "junior students," —our own impression is that anyone who could answer all these questions would be a well-rounded crystallographer indeed. A teacher looking for classroom exercises will spend many a stimulating evening browsing through these miscellanea. He will find that some of the problems are not easy; that others take time. The latter are more in the nature of seminar topics. For example, Problem No. 4 on page 111 is stated as follows: "Data: An edge of a crystallographic polyhedron can be seen (1) as the line of intersection of two faces; (2) as the axis of a zone; (3) as the line connecting two coigns of the polyhedron. Questions: (*a*) Study methods for the construction of crystallographic drawings in the following books and articles: A. E. H. Tutton . . . , M. Reeks . . . , A. Nies . . . , R. L. Parker . . . , M. W. Porter . . . , C. Viola . . . , L. Weber . . . , L. Burmester . . . , P. Terpstra . . . (*b*) Divide after the above quoted characteristics these methods in three groups." The answers to the questions are not given; we wish the author would publish them in a companion pamphlet.

The Table of Contents is missing. If there were one it would read as follows (number of pages in each chapter given in parentheses): I. Miller indices in the rhombohedral system (4), II. Geometrical crystallography (34), III. Rhombic section (4), IV. Transformation of co-ordinates (7), V. Twinning (10), VI. Gliding (23), VII. Lattices (20), VIII. Crystal drawing (8), IX. Optics (29), X. Laue patterns; Space groups; Weissenberg patterns (41), XI. Structure factors and Fourier series (8). A two-page index refers to subjects and to authors.

The order in which the chapters are arranged is somewhat puzzling, but their relative lengths seem well suited to give the student a thorough foundation in morphology and optics, and bring him up to structural crystallography.

After four books in Dutch, Professor Terpstra has decided to write this one in English. The going is rough in spots, but his readers will gladly grant him the leniency for which he begs in the preface. Long passages are quoted in either French or German; a Latin motto and its Dutch translation, on the last page, give a final polyglot touch. The "1001" questions are not numbered, we did not count them, but we bet you there are more!

Donnay and Donnay

FOURIERSYNTHESE VON KRISTALLEN UND IHRE ANWENDUNG IN DER CHEMIE,

by W. Nowacki. 238+13 pp., 120 figs., and 28 tables. Verlag Birkhäuser, Basel. Lehrbücher und Monographien aus dem Gebiete der exakten Wissenschaften, Chemische Reihe, Band VI, 1952. Cloth bound. Price Sw. Fr. 34.30 ($8.00).

This is the first comprehensive account of Fourier methods in structural crystallography. The author is professor of crystallography at the University of Bern, Switzerland. That such a monograph should appear in German is not surprising since the methods have been described almost exclusively in English publications—it does fill a special need for German workers. The foremost impression which the American reader receives is that of a well organized and thorough literature survey. It includes most papers up to 1948 and a few 1949 titles.

The first part is theoretical. It begins with Fourier series whose coefficients are structure factors, F. Electron-density sections as well as projections are treated. Here, as throughout the book, numerical examples from the literature are given in detail. Then follow discussions of accuracy of parameters, refinement methods, conversion of the data to an absolute scale, and phase determination. The second chapter covers $|F|^2$-syntheses of all kinds. The second part is practical. It deals with calculating aids: strips, punched-card machines, mechanical and other analogue computers. Special sections are devoted to optical methods of synthesis and the abortive attempt to use microwaves. Three alphabetical indexes are appended: authors, subject matters, and substances mentioned. Extinction tables appear in a separate, paper-bound, supplement; they list the presence criteria and the space groups corresponding to Buerger's 120 diffraction symbols (no alternate orientations are given).

As a textbook for advanced students, the treatment is not quite thorough enough. A considerable number of derivations are given, but many stumbling blocks are passed over, and such statements as "es ist anschaulich klar" and "es lässt sich leicht zeigen" abound. Before he can intelligently apply the methods described, a student will still have to study the literature. Excellent bibliographics, given at the beginning of each section, will make this easy for him. They will also help scientists in neighboring fields, who might wish to embark on a crystal-structure determination. For such readers, however, a critical appraisal of the methods presented would have been desirable.

Here are a few minor comments. Following an infelicitous British practice, Nowacki reserves the name "Fourier syntheses" to F-summations and uses other designations for $|F|^2$-summations, even though they too are Fourier series. Contrary to custom he advocates (p. 13) using "structure amplitude" for F and "structure factor" for $|F|^2$. His "ideal Realkristall" is our ideally imperfect crystal. The language is on the whole encouragingly simple and clear. Three recurring abbreviations baffled the reviewer, who begs to offer her interpretations: *u.W.* = unseres Wissens, so far as we know; *u.U.* = unter Umständen, under certain circumstances; *m.a.W.* = mit anderen Worten, in other words. The book has been reviewed by F. A. Bannister (*Min. Abs.* **11,** 513, Sept. 1952), and by D. P. Shoemaker (*Acta Cryst.*, **6,** 224, Feb. 1953).

Crystallographers will be indebted to the author for his useful compilation. All will share the hope that this volume can be kept up to date in future editions.

GABRIELLE DONNAY

ERZMIKROSKOPISCHES PRAKTIKUM, by HANS SCHNEIDERHÖHN. E. Schweizbart'sche Verlagsbuchhandlung (Erwin Nägele), Stuttgart. 1952. Price (bound in linen) DM 40.60.

This work of Professor Schneiderhöhn is essentially a new edition of the *Lehrbuch der Erzmikroskopie*, Vol. 1, Pt. I (Gebrüder Borntraeger) 1934, by Schneiderhöhn and Ramdohr. It will be remembered, perhaps, that the first part of this important "Lehrbuch" was largely the work of Schneiderhöhn. He has now revised it to include the new advances since 1934. The changes in the book are largely elimination and simplification. The section on microchemical analysis has been eliminated, and those regarding spectroscopical and x-ray examinations much reduced. The long and rather theoretical discussion by Doris Korn on the determination of the optical orientation of an opaque mineral in a polished section in the original volume has also been dropped. Otherwise the new book is little changed, even as to its wording.

Two chapters are new: Observations on Individual Minerals; and The Structure of

Aggregates. In the first named chapter general remarks are included concerning such topics as grain shape and size, twinning, zonal structure, intergrowths, exsolution structures, etc.; in the second, general descriptions of the various types of structures encountered in ore-mineral aggregates.

The text is supplemented by 32 plates with 64 figures of beautifully photographed and reproduced photomicrographs, most of which are new. A bibliography of 262 titles combines all the titles referred to in the text. A separate section of determinative tables occupies a pocket at the end of the volume. These tables include a determinative one of 3 pages, based upon the reflection capacity as measured by the microphotometer, and a second of 20 pages of the individual ore minerals, arranged alphabetically, giving the pertinent diagnostic optical, physical and x-ray data for 189 minerals.

This new work brings the highly useful Part I of the *Lehrbuch der Erzmikroskopie* up to date. Part II has already been revised by Ramdohr as *Die Erzmineralien und ihre Verwachsungen* (Akademie Verlag, Berlin. 1950). The printing and paper are good and the German text clear and easily comprehensible for the American reader. The binding, however, is flimsy for a book that will be submitted to frequent use, and the book and cover seem destined to part early in its life. This was done, perhaps, in an attempt to reduce the cost of the book to meet the constant complaint that scientific books and periodicals now printed in Germany are too high priced for many purchasers. The cost will certainly be a deterrent to its acquisition to many students, to which this work is especially directed. Nevertheless, the fact that there is no suitable substitute for this essential work should give it a wide market.

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DANA'S MANUAL OF MINERALOGY, sixteenth edition by Cornelius S. Hurlbut, Jr., viii+530 pp., 471 figures, XXII plates, frontispiece in color.

This sixteenth edition of *Dana's Manual of Mineralogy* is the second revision made by Professor Hurlbut and represents a far less sweeping revision than the first. The purpose of the revision seems to have been twofold: to bring the Manual abreast of developments in the field; to make the Manual pedagogically more sound. Both purposes have been achieved; the first by the addition of new data, the second by addition to and amendment of certain abstruse passages of the fifteenth edition. The new format is no particular improvement but is attractive and readable.

A new section on the *History of Mineralogy* has been added. In one section of this *History* Hurlbut states:

"During the latter part of the nineteenth century, mineralogy, for want of a tool sensitive and powerful enough to explore the fine structure of matter and affirm or deny the existence of Haüy's integral molecules, entered upon an unproductive period. . . . The situation was finally resolved in 1912 when x-rays . . . were demonstrated by Max von Laue to be the tool that would unlock the secrets of the crystal."

This sort of statement which labels the excellent and necessary descriptive work of the nineteenth and early twentieth centuries as *unproductive* is typical of the attitude of too many mineralogists today. If we examine Dr. Hurlbut's Manual we find that at least 90% of it could have been written prior to the twentieth century, and justly so. The advances of crystal chemistry have not negated the fact that mineralogy, as a subject, remains a natural science for the most part.

Professor Hurlbut has expanded his one page introduction of the fifteenth edition into a pedagogically improved seven pages entitled the *Nature of Minerals*. He has chosen in this edition not to define a *mineral* rigorously. (No mention is made of the fact that minerals are defined on the basis of their physical properties as well as their chemical properties.) He apparently now believes it is pedagogically better to derive concepts than to state them and explain them. From this reviewer's experience the other approach, which Hurlbut used in the fifteenth edition, has proved more satisfactory, in general.

Five pages are devoted to the new subject of *Applications of Mineralogy*, apparently to convince the student that the subject on which he is embarking is of adequate importance.

The new section on *Crystal Chemistry* represents the greatest departure from the fifteenth edition. Fourteen pages are devoted to such subjects as: *relation of chemistry to crystallography, polymorphism, structure, ionic bonds, covalent bond, van der Waal's bond, metallic bond, atomic packing, isomorphism, isostructure, exsolution, and homeomorphism.* Although the section is very clearly and very well written, the reader of the text is faced with the unhappy realization that there is very little application of the newly won principles in the other section of the book which is devoted to the treatment of individual minerals.

Another important and desirable difference between the fifteenth and sixteenth editions is the adoption in the latter of the Groth-Koksharow form nomenclature. Such terms as *macro, brachy, ortho, clino* have been deleted. Dr. Hurlbut has not seen fit to completely adopt the Groth Koksharow nomenclature but the exceptions have some pedagogical virtue.

A page is devoted to the meaning of the Hermann-Mauguin symmetry symbols, but no pedagogic use is made of them in the section on crystallography. They are not only good to know about; they are good to use as tools in the better understanding of crystal symmetries. On page 29, item 5 the positions of the symmetry elements for $\bar{6}m2$ are incorrectly stated. Crystal systems are defined in this revision on the basis of symmetry, rather than on relative lengths and angular relations of the axes.

The relative importance of the various mineral species is indicated in the text by the case of letters.

It is not possible to list here all of the new data which are included in the mineral descriptions. A few, however, will serve to demonstrate the painstaking care Dr. Hurlbut put into the revision of mineral descriptions, where necessary:

p. 210 Skutterudite replaces smaltite as a mineral name.

p. 230 New locality for emery; enlarged discussion on synthesis of corundum.

p. 261 New localities for mining of fluorite; mentions new use of optical fluorite.

p. 286 Places sulfates and chromates before instead of after phosphates *et al.*

p. 301 Adds (OH) to chemical formula of apatite.

p. 307 Corrects turquoise formula.

p. 309 Includes carnotite, which was not in fifteenth edition, but does not include it in mineral index in the rear of text.

p. 314 Enlarges bases for classification of silicates and gives figures as illustrations of each of the six silicate structural types. It is unfortunate that these structural type headings are not introduced into the proper places in the text of the actual descriptions of silicate species.

p. 374 Reassessment and restatement of occurrence of tourmaline.

p. 379 Lawsonite has been added, just why, it is hard to say.

p. 412 The beryl crystal from Albany, Maine, has apparently grown since the fifteenth edition.

p. 431 and following pages. More recent production and uses data for various substances.

In a completely reset text of this sort some typesetting errors will necessarily escape all proofreaders. A few of these plus some text errors plus some comments on sections of the text follow:

p. 8 Separates crystallography from physical mineralogy, but it is certainly a physical property.
p. 10 In the enlargement of various sections of the introduction, there has been little expansion of the *occurrence* of minerals under *descriptive* mineralogy. In a natural science such as mineralogy more attention to occurrence should be given in these introductory remarks. Parts of the discussion on pages 13, 14, and 15 belong under this heading.
p. 11 Not all *economic geologists* concentrate in the *ore minerals*.
p. 17 The new definition of cryptocrystalline is an improvement.
p. 18 Describes magmas as melts, rather than solutions.
p. 123 *Polar solvents* needs explanation or should be eliminated.
p. 124 The *middle* of the periodic table means nothing here. It is actually confusing, particularly since there are so many periodic tables extant.
p. 125 Line 3. "*Since tetrahedral groups, the basic units of structure*" should indicate that they are *basic for silicates*.
p. 129 Line 22. "It is becoming . . . clear that the unit of description (should often be, not *is*, reviewer) the mineral series."
p. 132 *This process of segregation and growth of rejected ions into autonomous crystal domains in the solid state from a disordered non equilibrium crystal is called exsolution*. What a sentence!
p. 250 Shouldn't limonite be placed in a category of a mixture as is bauxite?
p. 378 No mention of the recent great demand for beryl.
p. 401 There is nothing in the word *igneous* which implies a fluid origin. No mention is made of the fact that the formation of magmas is a subsurface phenomenon.
p. 402 No mention is made of the very important effect of volatiles on the texture of igneous rocks.
p. 414 Breccia and tillite should be mentioned. Marine conglomerates are ignored.
p. 415 No mention of flint or chert. The latter is certainly of considerable volumetric importance.

The above corrections and suggestions in no way detract from the over-all continued usefulness of this Manual as a textbook of mineralogy where it ranks, without question, amongst the most usable and pedagogically valuable books of its kind.

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TRANSACTIONS OF THE EDINBURGH GEOLOGICAL SOCIETY, Vol. XV, Campbell Volume. 408 pp. ed. by Arthur Holmes. 1942, 30s ($4.35).

This issue of the *Transactions of the Edinburgh Geological Society* is dedicated to Dr. Robert Campbell, who retired in September, 1951, after 47 years of devoted service as a teacher of mineralogy, petrology and geology, as a researcher, and as an administrator. The contributions by his students, associates and friends number 27 and are mainly in petrology, but papers on paleontology, glaciology, geochemistry, tectonics and geological education also are included. Many of the petrological papers deal with occurrences in Scotland. A high standard of quality is maintained throughout.

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NEW MINERAL NAMES

Yugawaralite

K. Sakurai and A. Hayashi, "Yugawaralite," a new zeolite. *Sci. Repts. Yokohama National Univ.*, Sec. II, No. 1, 69–77 (1952) (in English).

The mineral occurs in net works and veins and as crystals in cavities in andesite tuffs that have been altered by waters of hot springs. The locality is 300 m. from the Yugawara hot spring, Kanagawa Prefecture, Japan. The mineral is colorless to white, luster vitreous, *b* (010) commonly iridescent. Hardness $4\frac{1}{2}$, $G.=2.201$, 2.198. Monoclinic, flat crystals with *b* prominent up to 5 mm. (*a*-axis), 2 mm. (*b*-axis), 10 mm. (*c*-axis). Faces noted: *a* (100), *b* (010), *c* (001), *m* (110), *l* (120), *u* (011), *p* ($\bar{1}11$). Goniometric measurements gave $a:b:c=0.9758:1:0.7129$, $\beta=68°28'$. Cleavage imperfect on (010). *X*-ray oscillation photographs (MoKα) gave a_0 13.26, b_0 13.65, c_0 9.73 Å, β 68°30′. This gives $a:b:c=0.9715:1:0.7136$. *X*-ray powder data are given and compared with those for mordenite, epistilbite, heulandite, and stilbite. The strongest lines are (Å.) 5.69, 4.71, 3.72, and 3.07.

The mineral is optically positive, α—1.495, β—1.497, γ—1.504, 2V="56°, 68°, 76°, 89°, etc." on different specimens. Optical axial plane perpendicular to (010); extinction straight, elongation negative, absorption $r<v$.

Analyses by A. Hayashi and N. Simoda gave SiO_2 57.94, 58.44; Al_2O_3 17.65, 17.31; Fe_2O_3 0.35, 0.36; CaO 9.79, 9.75; MgO 0.86, 0.42; Na_2O 0.38, 0.36; K_2O 0.41, 0.12; H_2O^- 1.80, 2.03; H_2O^+ 10.70, 10.55; sum 99.88, 99.34%. These correspond to $Ca_4Al_7Si_{20}O_{54}\cdot 14H_2O$. The unit cell contains 1 molecule. Yugawaralite is not attacked by either cold or hot hydrochloric acid and does not gelatinize. A dehydration curve is given and compared with those given by laumontite, stilbite, heulandite, and ptilolite. About $3H_2O$ is lost to 100°, about $6H_2O$ from 100 to 400°, and about $4H_2O$ from 400 to 450°. A differential thermal analysis shows a marked endothermal peak at 540°. The name is for the locality.

Michael Fleischer

Zinc hoegbomite

V. A. Moleva and V. S. Miasnikov, Hoegbomite and its variety—zinc hoegbomite. *Doklady Akad. Nauk S.S.S.R.*, **83**, 733–736 (1952); from a translation kindly made by Mr. V. L. Skitsky.

Complete analyses and optical data are given for three samples of hoegbomite, but the localities are not stated. One sample was found in chlorite rock, associated with spinel, magnetite, epidote, and apatite. Analysis of the hoegbomite gave ZnO 11.12, MgO 9.07, FeO 8.41, MnO 0.80, Fe_2O_3 9.32, Cr_2O_3 none, Al_2O_3 56.67, TiO_2 5.12; sum 100.51%. This gives $MgO:ZnO:FeO=0.227:0.137:0.117$. This variety is called zinc hoegbomite. The other analyses show ZnO none, 1.12%.

Discussion: Should be called zincian hoegbomite.

M. F.